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# STRATIGRAPHY AND GEOLOGICAL CORRELATION

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**Boris S. Sokolov**

Member of the Russian Academy of Sciences
Counsellor of the Presidium of the Russian Academy of Sciences

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# Stratigraphy and Geological Correlation

*(Stratigrafiya, Geologicheskaya Korrelyatsiya)*

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# Regularities in the Late Cretaceous Flora Evolution in Northwestern Kamchatka and Koryak Highland

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**Abstract**—Processes responsible for the alternation of flora evolution stages are analyzed on the basis of the Late Cretaceous floras from northeastern Russia, among them two paleofloral successions from the Penzhina Bay (Turonian–Campanian) and the Rarytkin Range (middle Maastrichtian–Danian). Changes in the species diversity and ecosystem structures, as well as the extent of extinction and origin of taxons, mainly depend on the mode, trend, and sharpness of climatic changes at the stage boundaries and on the flora taxonomic composition. The flora renewal at each stage is significant (40–75% at the species level) and controlled by the intense evolution of flowering plants. However, there are many transient genera and species of conifers, which often become dominant or subdominant elements among various Late Cretaceous floras. Climatic conditions, especially humudification, affect the extent of geographic differentiation of floras as well.

**Key words:** *evolution of paleofloras, Late Cretaceous, northeastern Russia, climate*

Upper Cretaceous terrestrial deposits with plant fossils are widespread in northwestern Kamchatka and the Koryak Highland. In this region, it is possible to trace an almost continuous succession of floral assemblages ranging in age from the late Albian–Cenomanian to the Danian. The succession of taxonomically diverse and reliably dated floras may serve as the basis for studying the Late Cretaceous flora evolution in northeastern Russia. The taxonomy of most Late Cretaceous floras in the region has already been studied and described in detail (Herman and Lebedev, 1991; Shchepetov *et al.*, 1992; Golovneva, 1994). The main objective of this paper is to reveal the general regularities in the compositional changes of floral assemblages at the boundaries dividing their evolutionary stages. The names of all the plants mentioned in the paper are given without revision and in accord with the monographs devoted to concrete floras. We used several successions of paleofloras from the Koryak Highland and northwestern Kamchatka as a basis for the analysis.

A continuous succession of floral assemblages ranging in age from the Turonian to Campanian (Herman and Lebedev, 1991) can be observed in several adjacent sections of the Konglomeratovyi Cape, Valizhgen Cape, and Elistratov Peninsula of the Penzhina Bay (Fig. 1). These are the Penzhina (late Turonian), Kaivayam (Coniacian), Valizhgen (possibly early Santonian) and Upper Bystraya (early Campanian) assemblages (Table 1). Fossil plants are recovered from deposits of the Valizhgen and Bystraya formations interrelated through a gradual lithological transition.

The second continuous succession of floral assemblages is detected in deposits of the Rarytkin Formation of the Gornaya River basin (northeastern part of the Rarytkin Range). It comprises two assemblages (Golovneva, 1994): the middle Maastrichtian Gornaya and the late Maastrichtian–Danian Rarytkin floras (Table 2).

In addition, we considered the Grebenka assemblage (late Albian–Cenomanian to early Turonian) from the Krivorechenskaya Formation of the Anadyr River middle course, the Ginter (middle Cenomanian) and Barykov (early Campanian) assemblages from synonymous formations of the Ugolnaya Bay, the Kakanaut assemblage (middle Maastrichtian) from the Kakanaut Formation of the Kakanaut River basin, and the Koryak assemblage (late Maastrichtian) from the Koryak Formation in the Emima River basin (Fig. 1).

Throughout the Cretaceous, these sites apparently were situated at the periphery of a marine basin. Alluvial, lacustrine, marshy, often coal-bearing deposits, in which the plant remains were buried, accumulated on the marine plain. Uplands of the Okhotsk–Chukotka volcanogenic belt were arranged northwest of the plain, and, therefore, terrigenous rocks often contain tuffaceous material. Continental sedimentation was intermittently replaced by marine. The alternation and partial substitution of marine sediments by continental facies is favorable for the precise dating of floral assemblages.

The Late Cretaceous evolution of flora in northeastern Russia can be subdivided into the Grebenka (Cenomanian–early Turonian), Penzhina (late Turonian),

**Table 1.** Distribution of species in the late Turonian–early Campanian floral assemblages

| Penzhina | Kaivayam | Valizhgen | Verkhnyaya Bystraya and Barykov | Species |
|---|---|---|---|---|
| late Turonian | Coniacian | early Santonian | early Campanian | |
| | | | | *Sphenopteris crispans* E. Lebedev |
| | | | | *Cladophlebidium singulinervis* E. Lebedev |
| | | | | *Cladophlebis contrarius* E. Lebedev |
| | | | | *Cladophlebis* sp. 1 |
| | | | | *Cladophlebis* sp. 2 |
| | | | | *Protophyllocladus sachalinensis* (Krysht. et Baik.) Krassil. |
| | | | | *Paraprotophyllum cordatum* (Samyl.) Samyl. |
| | | | | *Dicotylophyllum* sp. 1 |
| | | | | *Grebenkia* aff. *kryshtofovichii* E. Lebedev |
| | | | | *Penzhinia vachrameevii* Herman |
| | | | | *Celastrophyllum* aff. *kolymensis* Samyl. |
| | | | | *C. acutidens* Fontaine |
| | | | | *Arthollia inordinata* (Hollick) Herman |
| | | | | *Platanus primaeva* Lesquereux |
| | | | | *Menispermites septentrionalis* Hollick |
| | | | | *Celastrophyllum retinerve* Herman |
| | | | | *Platanus newberryana* Heer |
| | | | | *Arthollia pacifica* Herman |
| | | | | *Myrtophyllum penzhinense* Herman |
| | | | | *Trochodendroides notabilis* Herman |
| | | | | *Gleichenia pseudocrenata* E. Lebedev |
| | | | | *Onychiopsis grandipinnula* E. Lebedev |
| | | | | *Menispermites kryshtofovichii* Vachrameev |
| | | | | *Cissites inscissus* Herman |
| | | | | *Dalembia pergamentii* Herman et Lebedev |
| | | | | *Viburniphyllum whymperi* (Heer) Herman |
| | | | | *Sequoia teniufolia* (Schmalh.) Sveshn. et Budants. |
| | | | | *Asplenium* ex gr. *dicksonianum* Heer |
| | | | | *Pityophyllum* ex gr. *nordenskioldii* (Heer) Nathorst |
| | | | | *"Zizyphus" smilacifolia* Budants. |
| | | | | *Paraprotophyllum ignatianum* (Krysht. et Baik.) Herman |
| | | | | *Marchantites jimboi* (Krysht.) Krysht. |
| | | | | *Arctopteris penzhinensis* E. Lebedev |
| | | | | *"Cephalotaxopsis" heterophylla* Hollick |
| | | | | *Ginkgo* ex gr. *adiantoides* (Unger) Heer |
| | | | | *Paraprotophyllum pseudopeltatum* Herman |
| | | | | *Arthollia insignis* Herman |
| | | | | *Magnoliaephyllum alternans* (Heer) Seward |
| | | | | *Dicotylophyllum dombeyopsoidum* Herman |
| | | | | *Ochotopteris kamtchatica* E. Lebedev |
| | | | | *Birisia jelisejevii* (Krysht.) Philippova |
| | | | | *B. oerstedtii* (Heer) E. Lebedev |
| | | | | *Ternstroemites harwoodensis* (Dawson) Bell |
| | | | | *Sapotacites* aff. *alaskensis* Hollick |
| | | | | *Araliaephyllum leonovii* Herman |
| | | | | *A. quinquelobum* Herman |
| | | | | *Cissites asymmetricus* Herman |

**Table 1.** (Contd.)

| Penzhina | Kaivayam | Valizhgen | Verkhnyaya Bystraya and Barykov | Species |
|---|---|---|---|---|
| late Turonian | Coniacian | early Santonian | early Campanian | |
| | | | | *Dicotylophyllum longifoluim* (Krysht. et Baik.) Herman |
| | | | | *D. longipetiolatum* Herman |
| | | | | *Dicotylophyllum* sp. 2 |
| | | | | *Osmunda* sp. |
| | | | | *Ruffordia magnifolia* E. Lebedev |
| | | | | *Dalembia krassilovii* Herman et Lebedev |
| | | | | *Platanus prisca* Herman |
| | | | | *Equisetites* sp. |
| | | | | *Elatocladus smittiana* (Heer) Seward |
| | | | | *Metasequoia cuneata* (Newb.) Chaney |
| | | | | *Cupressinocladus cretaceus* (Heer) Seward |
| | | | | *Magnoliaephyllum magnificum* (Dawson) Krassilov |
| | | | | *Viburniphyllum lebedevii* Vachr. et Herman |
| | | | | *Quereuxia angulata* (Newb.) Krysht. |
| | | | | *Anemia* sp. |
| | | | | *Thallites* sp. |
| | | | | *Sagenopteris* sp. |
| | | | | *Cycadites* aff. *hyperborea* (Krysht.) E. Lebedev |
| | | | | *Pityostrobus* sp. |
| | | | | *Carpolithes* sp. |
| | | | | *Coniopteridium* sp. |
| | | | | *Hausmannia* sp. |
| | | | | *Cladophlebis venustus* E. Lebedev |
| | | | | *Cladophlebis* sp. 2 |
| | | | | *Macclintockia crenata* Herman |
| | | | | *Sequoia* aff. *reichenbachii* (Gein.) Heer |
| | | | | *Nilssonia* aff. *alaskana* Heer |
| | | | | *Viburniphyllum basicordatum* Herman |
| | | | | *Celastrinites septentrionalis* (Krysht.) Golovn. |
| | | | | *Menispermites* sp. |
| | | | | "*Grewiopsis*" *akhmetjevii* Vachr. et Herman |
| | | | | "*Quercus*" *tchucotica* Abramova |
| | | | | *Macclintockia ochotica* Vachr. et Herman |
| | | | | *Penzhinopteris pergamentii* E. Lebedev |
| | | | | *Nilssonia* aff. *yukonensis* Hollick |
| | | | | *Pterophyllum validum* Hollick |
| | | | | "*Vitis*" *penzhinica* Herman |
| | | | | *Cissites beljaevii* Herman |
| | | | | *Viburniphyllum alnoides* (Hollick) Herman |
| | | | | *Sequoia fastigiata* (Sternb.) Heer |
| | | | | *Nymphaeites* sp. |
| | | | | *Platanus* cf. *marginata* (Lesq.) Heer |
| | | | | "*Credneria*" sp. |
| | | | | *Araliaephyllum polevoii* (Krysht.) Krassilov |
| | | | | "*Vitis*" *stantoni* (Knowlton) Brown |
| | | | | *Trochodendroides arctica* (Heer) Berry |
| 35 | 45 | 29 | 35 | Number of species |

▬▬ Dominant species; ▬ Representative species; — Rare species.

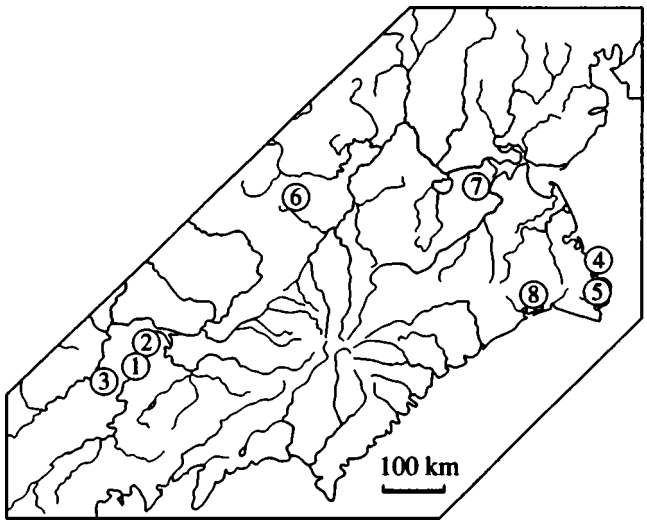


**Fig. 1.** Principal localities of the Late Cretaceous fossil floras in northeastern Russia: (1) Konglomeratovyi Cape; (2) Valizhgen Cape; (3) Elistratov Peninsula; (4) Ugolnaya Bay; (5) Amaam Lagoon and basins of the Emima and Ilnaivaam rivers; (6) Grebenka River; (7) Rarytkin Range; (8) Kakanaut River basin.

Kaivayam (Coniacian), Barykov (Santonian–early Campanian), Gornaya (the middle Maastrichtian), and Rarytkin or Koryak (late Maastrichtian–Danian) regional stages. Correlation of floral assemblages with the flora evolution stages is shown in Table 3.

The Grebenka flora is first in a series of Cretaceous floras of NE Russia displaying a predominance of angiosperms. The angiosperm species constitute 35–40% of its composition. They are followed by conifers (25%), ferns (23%), and by other groups of plants (Table 4, Fig. 2). The evolution of the flora falls on the thermal maximum. Owing to this, the flora is characterized by a high taxonomic diversity and includes more than 80 plant species (Shchepetov *et al.*, 1992). As far as the taxonomy is concerned, the Grebenka flora comprises both the older Early Cretaceous and younger Late Cretaceous taxons. Ferns of the *Coniopteris* genus are most abundant in the flora; *Birisia*, *Hausmannia* and *Cladophlebis* genera are also typical taxons. *Gleichenia*, *Arctopteris*, *Asplenium* and *Sphenopteris* genera are rare. These fern taxons are typical of floras of the latter half of the Early Cretaceous. Abundant cycadophytes in the Grebenka flora are indicative of the warm climate. They are represented by *Taeniopteris*, *Cycadites hyperborea* (Krysht.) E. Lebed., some species of the *Nilssonia* genus, and by a new species of the *Nilssoniocladus* genus. Fragments of *Cycadites* and *Nilssonia* usually occur in oryctocenoses in the form of monodominant accumulations. Ginkgoaleans are common. Representatives of *Ginkgo* ex gr. *lepida* Heer with more divided leaves typical of older forms, *Baiera*, and *Sphenobaiera* were found together with *Gingko* ex gr. *adiantoides* (Unger) Heer. Czekanowskialeans are represented by individual imprints of *Phoenicopsis*. They become relict in the Grebenka flora. "*Cephalotaxopsis*," *Araucarites* and *Elatocladus* genera of conifers, which are principal components of most Late Cretaceous floras, are abundant. Fragments of *Sequoia* and various Pinaceae (*Pityophyllum*, *Pityostrobus*, *Pityospermum*, *Pseudolarix*) are rare. Representatives of *Athrotaxopsis*, *Pagiophyllum*, and *Podozamites*, the relict taxon for this flora, are more rare. Various angiosperm species of the *Menispermites* genus are abundant, and large-leaved Platanaceae together with representatives of *Magnoliaephyllum*, *Dalbergites*, *Scheffleraephyllum*, *Grebenkia*, *Celastrophyllum*, *Cissites*, *Dalembia*, and *Sorbites* are common. Representatives of the *Trochodendroides* genus are scarce. Many plants of the Grebenka flora had entire leaves, which also confirms the thermophilic affinity of this flora. The abundance of large-leaved Platanaceae in some occurrences testifies that the climate of the Grebenka time was very warm and humid. It should be pointed out that Platanaceae were widespread along river banks at that time and did not colonize other areas. Imprints of their leaves are mainly confined to the river-bed facies.

The climate became cooler in the late Turonian. The species diversity of the Penzhina stage was reduced nearly by one half (Table 4). All taxonomic groups of plants experienced reduction because of extinction and migration of some thermophilic species in a southward direction. Some species and genera of the Grebenka flora missing in the Penzhina assemblage reappeared in northeastern Russia as components of the Kaivayam and younger assemblages in response to the successive warming of climate. Among forms that migrated southwards, there were primarily cycadophytes missing in the Penzhina assemblage, then angiosperms with entire leaves (genera *Magnoliaephyllum*, *Araliaephyllum*, *Cissites*), some thermophilic ferns, and conifers. Extinction at this stage was substantial (Table 5). Most Early Cretaceous plants, which played a significant part or became relict in the Grebenka flora, disappeared. These are old conifers (*Podozamites*, *Pagiophyllum*, *Athrotaxopsis*), czekanowskialeans (*Phoenicopsis*), some ginkgoaleans (*Ginkgo* ex gr. *lepida*, *Sphenobaiera*, *Baiera*), some fern taxa (first of all, the diverse *Coniopteris* and some species of *Birisia*, *Gleichenia*, and *Cladophlebis*). Small-leaved taxons of angiosperms, which existed in various Early Cretaceous floras, became extinct. These are *Dalbergites*, *Leguminosites*, some *Celastrophyllum* forms, and *Cissites*, as well as the diverse thermophile taxons, which failed to migrate southward. Large-leaved, cold-resistant Platanaceae forms, *Menispermites*, and some species of *Dalembia* and *Grebenkia* passed from the Grebenka to the Penzhina flora.

Angiosperms, mainly the large-leaved Platanaceae, dominate in the Penzhina flora (57% as compared to 35% in the Grebenka assemblage). Platanaceae seem to reach the greatest taxonomic and ecological diversity in the late Turonian floras. Their fossil leaves are found

**Table 2.** Distribution of species (the middle Maastrichtian–Danian) in floral assemblages of the Rarytkin Formation

| Cretaceous | | | Paleogene | Species |
|---|---|---|---|---|
| Maastrichtian | | | Danian | |
| middle | | late | | |
| Gornaya | | Rarytkin | | |
| early | late | early | late | |
| | | | | *Osmunda* sp. |
| | | | | *Cryptomerites ancistrophyllus* Golovn. |
| | | | | *Nuphar* sp. |
| | | | | *Trochodendroides longifolia* Golovn. |
| | | | | *Quereuxia flabellata* Golovn. |
| | | | | *Q. rotundifolia* Golovn. |
| | | | | *Palaeotrapa triangulata* Golovn. |
| | | | | *Viburnum stenophyllum* Vassilevsk. et Golovn. |
| | | | | *Cissus pseudomarginata* Golovn. |
| | | | | *Rhizomites* sp. B. |
| | | | | *Nilssonia serotina* Heer |
| | | | | *Sequoia minuta* Sveshn. |
| | | | | *Parataxodium* sp. |
| | | | | *Trochodendroides bifida* Golovn. |
| | | | | *Pecilnea pinnatiloba* Golovn. |
| | | | | *Carpolithes ceratops* Golovn. |
| | | | | *C. ellipticus* Golovn. |
| | | | | *Palaeotrapa bicornata* Golovn. |
| | | | | *Platimelis umcuvensis* Golovn. |
| | | | | *Arthollia inundata* Golovn. |
| | | | | *Viburnum oblongum* Golovn. |
| | | | | *Arctoterum rubifolium* Golovn. |
| | | | | *Platanus rarinervis* Golovn. |
| | | | | *Dyrana orientalis* Golovn. |
| | | | | *Renea actinidifolia* Golovn. |
| | | | | *Dicotylophyllum dentatum* Golovn. |
| | | | | *Carpolithes aculeatus* Golovn. |
| | | | | *Limnobiophyllum acutatum* (Dawson) Krassilov |
| | | | | *Carpolithes bulbus* Golovn. |
| | | | | *Rhizomites* sp. |
| | | | | *Viburnum acutifolium* Golovn. |
| | | | | *Sequoia* sp. |
| | | | | *Taxites* ex gr. *olrikii* Heer |
| | | | | *Equisetum arcticum* Heer |
| | | | | *Ginkgo* ex gr. *adiantoides* (Unger) Heer |
| | | | | *Metasequoia occidentalis* (Newb.) Chaney |
| | | | | *Microconium beringianum* Golovn. |
| | | | | *Glyptostrobus nordenskioldii* (Heer) R.W. Brown |
| | | | | *Trochodendroides arctica* (Heer) Berry |
| | | | | *Nyssidium arcticum* (Heer) Iljinskaya |
| | | | | *Corylus beringiana* (Krysht.) Golovn. |
| | | | | *Quereuxia angulata* (Newb.) Krysht. |
| | | | | *Celastrinites septentrionalis* (Krysht.) Golovn. |

**Table 2.** (Contd.)

| Cretaceous | | | Paleogene | Species |
|---|---|---|---|---|
| Maastrichtian | | | Danian | |
| middle | | late | | |
| Gornaya | | Rarytkin | | |
| early | late | early | late | |
| | | | | *Araucarites conophyllus* Golovn. |
| | | | | *Elatocladus talensis* Golovn. |
| | | | | *Arthollia rarytkensis* Golovn. |
| | | | | *Nyssa tshucotica* Golovn. |
| | | | | *Liriophyllum aeternum* Golovn. |
| | | | | *Viburnum rarytkense* Golovn. |
| | | | | *Dicotylophyllum acutidentatum* Golovn. |
| | | | | *D. incertum* Golovn. |
| | | | | *Rarytkinia terechovae* Vassilevsk. et Golovn. |
| | | | | *Haemanthophyllum cordatum* Golovn. |
| | | | | *Platanus raynoldsii* Newb. |
| | | | | *Disanthophyllum longipetiolatum* Vassilevsk. et Golovn. |
| | | | | *Quercus groenlandica* Heer |
| | | | | *Vitis olrikii* Heer |
| | | | | *Palaeotrapa aculeata* Golovn. |
| | | | | *Viburnum pseudoantiquum* Golovn. |
| | | | | *Coniopteris tschuktschorum* (Krysht.) Samyl. |
| | | | | *Trochodendroides bidentata* Vassilevsk. et Golovn. |
| | | | | *Platimelis platanoides* Golovn. |
| | | | | *Onoclea hesperia* R.W. Brown |
| | | | | *Taxodium* sp. |
| | | | | *Trochodendroides lamutensis* Golovn. |
| | | | | *Pseudolarix arctica* Krysht. |
| | | | | *Pityophyllum pseudolaricis* Krysht. |
| | | | | *Pityospermum* sp. |
| | | | | *Taxites cephalotaxifolius* Golovn. |
| | | | | *Ilex vernicosa* Golovn. |
| | | | | *Araliaephyllum pentalobatum* Golovn. |
| | | | | *Vitis* sp. |
| | | | | *Cissus* sp. |
| | | | | *Phragmites* ex gr. *alaskana* Heer |
| | | | | *Sparganiophyllum multinervosum* Golovn. |
| | | | | *Dicotylophyllum cercidifolium* Golovn. |
| | | | | *D. myricifolium* Golovn. |
| | | | | *D. acerifolium* Golovn. |
| | | | | *D. integerrimum* Golovn. |
| | | | | *Dicotylophyllum* sp. A |
| | | | | *Dicotylophyllum* sp. B |
| | | | | *Carpolithes alatus* Golovn. |
| | | | | *C. biloculatus* Golovn. |
| | | | | *Carpolithes* sp. A |
| | | | | *Carpolithes* sp. B |
| 35 | 33 | 36 | 48 | Number of species |

Dominant species; Representative species; Rare species.

**Table 3.** Stages of flora evolution and related principal floral assemblages of NE Russia

| Age (Odin, Kennedy, 1982) | | Regional stages of flora evolution | Anadyr River middle course | Northwestern Kamchatka | Ugolnaya Bay | Pekulnei Lake | Rarytkin Range |
|---|---|---|---|---|---|---|---|
| Ma | stage | | | | | | |
| −65 | Danian | Rarytkin | | | | | Rarytkin |
| | | | | | Koryak | | |
| | Maastrichtian | Gornaya | | | | Kakanaut | Gornaya |
| −72 | | | | | | | |
| | Campanian | Barykov | | Verkhnyaya Bystraya | Barykov | | |
| −83 | | | | | | | |
| | Santonian | | | Valizhgen | | | |
| −86 | Coniacian | Kaivayam | | Kaivayam | | | |
| −88 | | Penzhina | | Penzhina | | | |
| | Turonian | | | | | | |
| −91 | | | | | | | |
| | Cenomanian | Grebenka | Grebenka | | Ginter | | |
| −95 | | | | | | | |

not only in fluviatile sandstone beds, but also in finer-grained, often coal-bearing rocks, which accumulated in oxbows, lakes and swamps. In the Penzhina flora, Platanaceae are represented by genera *Platanus*, *Credneria*, *Paraprotophyllum*, and *Arthollia*. Representatives of the *Pseudoprotophyllum* genus had a dominant role in northerly floras of that time (Novosibirskie Islands, Alaska). Most species of the late Turonian Platanaceae are characterized by large leaves, great morphologic diversity, and wide geographic ranges. Extended geographic and significant vertical ranges are characteristic features for many plants of the Penzhina assemblage. These are low-specialized transient taxons occurring in the Cenomanian and sometimes even Albian temperate floras of Siberia, Alaska, and Kazakhstan: *Menispermites* ex gr. *septentrionalis*, *M. kryshtofovichii*, *Platanus newberryana*, *P. primaeva*, *Celastrophyllum kolymensis*, and *C. acutidens*. In addition to this "remanent" element of older floras, the locally formed taxons played an important part as well. *Paraprotophyllum ignatianum* and *Arthollia pacifica* dominate in the Penzhina assemblage. The former is a representative species of the Coniacian floras of Sakhalin and northeastern Russia, the latter is a local form. In addition to Platanaceae, the *Trochodendroides* genus played a leading role in the Penzhina assemblage. The endemic *T. notabilis* and *Zizyphus smilacifolia* are subdominant elements. The latter, like *Paraprotophyllum ignatianum*, is a representative species of the Coniacian floras of northeastern Russia and Siberia. Representatives of the *Trochodendroides* genus are rare in the pre-Turonian deposits. In addition to the plants mentioned above, *Myrtophyllum penzhinense* and *Celastrophyllum retinerve* are dominant species in the Penzhina assemblage. *Penzhinia vachrameevii* and *Dalembia pergamentii* are also typical species. The *Sequoia tenuifolia* and "*Cephalotaxopsis*" *heterophylla* species dominate among conifers. *Pityophyllum* ex gr. *nordenskioldii* is more rare. The genus *Protophyllocladus*, which is missing in the Cenomanian but typical of the Turonian floras, appears. Ferns are abundant and diverse. They are represented by nine species from genera *Sphenopteris*, *Cladophlebidium*, *Cladophlebis*, *Gleichenia*, *Onychiopsis*, *Asplenium*, and *Arctopteris*, which are characteristic of the Middle Cretaceous floras.

Large-leaved Platanaceae and representatives of the *Trochodendroides* genus dominate in the Kaivayam assemblage, like in the Penzhina one. In addition to them, a great number of thermophilic taxons usually having entire leaves appear. Among these are *Magnoliaephyllum alternans*, *M. magnificum*, *Ternstroemites hardwoodensis*, *Sapotacites*, and some species of *Araliaephyllum* and *Cissites* genera. Many of the thermophilic taxons migrated southwards during the late Turonian cooling to reappear in the Coniacian epoch of

**Table 4.** Abundance of taxonomic groups (%) in the Late Cretaceous floras of NE Russia

| | Floral assemblages | | | | | | |
|---|---|---|---|---|---|---|---|
| | Grebenka, Ginter | Penzhina | Kaivayam | Valizhgen | Verkhnyaya Bystraya, Barykov | Kakanaut, Gornaya | Rarytkin, Koryak |
| Mosses | 2.4 | 2.8 | 2.2 | 3.4 | 2.8 | 0 | 0 |
| Horsetails | 1.2 | 0 | 2.2 | 0 | 2.8 | 1.7 | 1.5 |
| Ferns | 23 | 25.7 | 20 | 24.1 | 8.5 | 5 | 6 |
| Cycadophytes | 7.2 | 0 | 0 | 6.9 | 8.5 | 3.4 | 0 |
| Gingkoaleans | 4.8 | 2.8 | 2.2 | 3.4 | 2.8 | 1.7 | 1.5 |
| Caytonialeans | 1.2 | 0 | 0 | 3.4 | 0 | 0 | 0 |
| Conifers | 25 | 11.2 | 13.3 | 20.7 | 20 | 13.5 | 17.9 |
| Flowering plants | 32 | 57.1 | 60 | 37.9 | 57.1 | 74 | 73 |
| Number of species | >80 | 35 | 45 | 29 | 35 | 59 | 67 |

warming, although, in the Kaivayam flora, many genera are represented by different species than in the Cenomanian.

When passing from the Penzhina to the Kaivayam assemblage, the Platanaceae taxons change only slightly. The changes occur at the species level. Whereas *Platanus primaeva* and *Paraprotophyllum cordatum* die out, *Platanus prisca* and *Paraprotophyllum pseudopeltatum* appear. Leaves of *Arthollia pacifica*, *Paraprophyllum ignatianum* and *Platanus newberryana* become less abundant. A new species, *Arthollia insignis*, appears to become dominant. Leaves of the *Viburniphyllum* genera, which are similar in morphology to those of Platanaceae, are common.

The *Trochodendroides* genus is represented by the same species that are present in the Penzhina assemblage, but the abundance of *Trochodendroides notabilis* is reduced while that of *Ziziphus smilacifolia* increases.

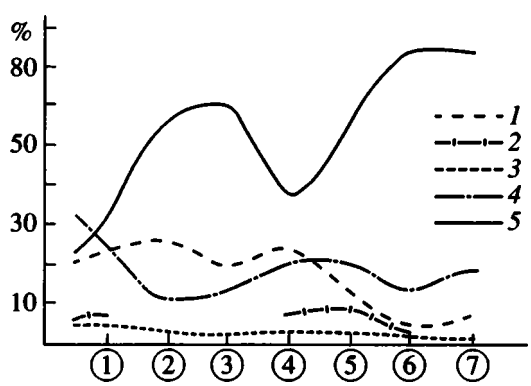


**Fig. 2.** Variations in abundance of main taxonomic plant groups in the Cretaceous floras of northeastern Russia: (*1*) ferns; (*2*) cycadophytes; (*3*) gingkoaleans; (*4*) conifers; (*5*) flowering plants. Numerals in circles indicate fossil floras: 1, the Grebenka; 2, Penzhina; 3, Kaivayam; 4, Valizhgen; 5, Barykov and Verkhnyaya Bystraya; 6, Gornaya and Kakanaut; 7, Rarytkin and Koryak.

The Kaivayam flora is characterized by a higher species diversity as compared to the Penzhina flora, though not exceeding the level of diversity of the Grebenka flora. An increase in the amount of angiosperms species with entire leaves, along with an increase in the species diversity of the flora as a whole, points to the climate warming at the Kaivayam time. The diversity and abundance of Platanaceae, which are the principal indications of humidity, change to only a small extent throughout the Cenomanian–Coniacian. This suggests that humidity was quite stable in this period.

Dominant species in the Kaivayam assemblage are *Paraprotophyllum pseudopeltatum*, *Arthollia insignis*, *Ziziphus smilacifolia*, *Viburniphyllum lebedevii*, and *Magnoliaephyllum alternans*. The *Arthollia pacifica*, *Myrtophyllum penzhinense*, *Trochodendroides notabilis*, *Dalembia pergamentii*, *Viburniphyllum whymperi*, *Paraprotophyllum ignatianum*, *Araliaephyllum leonovii*, *Magnoliaephyllum magnificum*, and some species of the *Dicotylophyllum* genus are also typical of the assemblage.

Ferns change moderately. Some older species of the *Cladophlebis* genus disappear. There appear more southern elements, part of which were still representative of the Grebenka flora: *Birisia jelisejevii*, *B. oertedtii*, *Osmunda* sp., *Ruffordia magnifolia* and some endemic species of the *Ochotopteris* genus.

*Sequoia tenuifolia* and "*Cephalotaxopsis*" *heterophylla* species, as well as *Pityophyllum* ex gr. *nordenskioldii* and *Elatocladus smittiana*, continue to dominate among conifers. Protophyllocladus sachalinensis becomes extinct. The appearance (at this stage) of a group of younger conifers from recent genera *Metasequoia cuneata* and *Thuja cretacea*, which are still rare in this flora, is of interest. Their abundance gradually increases in younger floras. The first appearance of the recent genera *Metasequoia*, *Glyptostrobus*, and *Thuja* of the Penzhina time is recorded not in the Penzhina

**Table 5.** Diversity of the Late Cretaceous floral assemblages in NE Russia

| | Floral assemblages | | | | | | |
|---|---|---|---|---|---|---|---|
| | Grebenka, Ginter | Penzhina | Kaivayam | Valizhgen | Verkhnyaya Bystraya, Barykov | Gornaya, Kakanaut | Rarytkin, Koryak |
| Total diversity | >80 | 35 | 45 | 29 | 35 | 59 | 67 |
| Amount of newly appeared species in the assemblage | | 20(57.1%) | 26(57.7%) | 17(58.6%) | 23(65.7%) | 45(67%) | 49(75%) |
| Amount of disappeared species in the assemblage | 65(81.3%) | 16(45.7%) | 33(73.3%) | 17(58.6%) | 21(60%) | 41(62%) | |
| Amount of species inherited from the preceding assemblage | | 15(42.8%) | 19(42.2%) | 12(41.3%) | 12(34.2%) | 14(23%) | 17(25%) |

assemblage, but in the coeval Volch'ya assemblage (Filippova, 1984) within the Okhotsk–Chukchi volcanogenic belt. The Kaivayam assemblage as a whole is successive to some extent with respect to the Penzhina and older Grebenka assemblages. Most species and genera of the Kaivayam assemblage, including dominant ones, originated from the Penzhina assemblage, only the abundance changing. The amounts of newly appeared species and the number of extinct taxons are insignificant. The changes occur at the species level. In addition to the inherited, there are many species in the Kaivayam assemblage, which migrated from the south.

Some older Cenomanian (representatives of *Celastrophyllum* and *Grebenkia*) and specific late Turonian taxonomic groups (*Penzhinia*) disappeared at the boundary between the Penzhina and Kaivayam assemblages. The propagation of younger conifers from recent genera *Metasequoia*, *Glyptostrobus*, *Thuja*, and *Pseudolarix* is the most remarkable event of that time.

Although the climatic warming was still in progress, the extinction of species and reduction in their diversity took place in the Valizhgen time (the early Santonian). A rise in the temperature is confirmed by the reappearance of thermophilic cycadophytes in northern floras, where they are abundant. The reduction of species diversity was due to a lower humidity, as is evident from a lesser diversity and abundance of ferns, the prevalence of hook-leaved conifers, and the disappearance of large-leaved Platanaceae. The only species *Platanus prisca* with small leaves has survived. Abundant thermophilic forms with entire leaves (*Magnoliaephyllum*, *Araliaephyllum*, and *Cissites*) became extinct as well. These forms seem to be adapted to a more humid climate. The lesser humidity dramatically affected the flora that evolved earlier under stable humid conditions for a long period of time, although this event was not very dramatic because coal accumulation on the coastal plains of northeastern Russia progressed throughout the Late Cretaceous.

The *Macclintockia crenata* species of angiosperms and coniferous *Sequoia* ex gr. *reichenbachii* (probably a new independent species with large, up to 1.5 cm, hook-like leaves) predominate in the Valizhgen assemblage. Gingkoaleans (*Ginkgo* ex gr. *adiantoides*) and cycadophytes, which are represented by the same genera and certain species of the Grebenka assemblage (*Cycadites hyperborea*) and two *Nilssonia* species, are abundant as well. The *Sagenopteris* genus reappears along with cycadophytes. Conifers characteristic of the Turonian–Coniacian floras (*Sequoia tenuifolia*, "*Cephalotaxopsis*," and diverse *Pinaceae*) become rare, whereas species of recent genera *Metasequoia*, *Glyptostrobus*, and *Thuja* disappear. The less diverse and abundant ferns experienced changes in the species composition. The typical Middle Cretaceous genera *Birisia*, *Onychiopsis*, *Gleichenia*, and *Ruffordia* vanish. The *Anemia*, *Hausmannia*, and *Coniopteridium* genera and some endemic species of *Cladophlebis* appear in the flora.

The temperature still was high in the early Campanian (the Barykovian regional stage), though the aridity of the climate slightly decreased.

*Macclintockia ochotica* (similar to *M. crenata* from the Valizhgen assemblage) and *Quercus tchucotica* dominate in the flora. Conifers are represented by *Sequoia fastigiata*, *Elatocladus smittiana*, "*Cephalotaxopsis*" *heterophylla*, and *Pityophyllum* sp. *Metasequoia*. *Thuja* genera reappear. Rare hydrophilous forms of Platanaceae (*Paraprotophyllum ignatianum*, *Platanus marginata*, *Credneria* sp.), *Menispermites* sp., and diverse species of *Viburniphyllum* appear. Representatives of riparian plants (*Equisetites* sp., *Marshantites jimboi*, *Quereuxia angulata*, *Nymphaeites* sp.) unknown in the Valizhgen time are common. Many new genera and species emerge: *Celastrophyllum septentrionalis*, *Grewiopsis akhmetjevii*, *Vitis penzhinica*, *V. stantonii*, *Cissites beljaevii*, and others. The proportion of forms of the *Trochodendroides* genus increases. Like in the Valizhgen assemblage, cycadophytes from genera *Nilssonia* and *Pterophyllum* are abundant. Ferns drastically reduce their diversity and abundance. They are represented by *Asplenium*, *Arctopteris*, and *Penzhinopteris*. A decrease in the abundance of ferns can not be explained by the climate aridity. It is likely that they

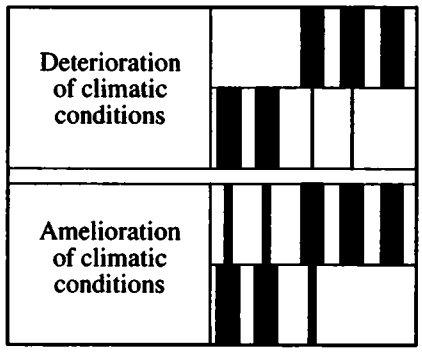


**Fig. 3.** Behavior of dominant species at the boundaries between stages in response to climatic changes.

were displaced from former habitat areas by flowering plants.

The Gornaya flora of the middle Maastrichtian follows the Barykov flora. No fossil floras have been found in the upper Campanian and lower Maastrichtian deposits of northeastern Russia. Accordingly, it is unclear whether or not the Gornaya flora is a successor of the Barykov flora, or whether any other flora has existed between them. An individual floral stage is suggested by the fact that a substantial cooling stage is assumed to fall at the boundary between the Campanian and Maastrichtian (Zakharov *et al.*, 1984). The Gornaya and Barykov floras are similar in general aspects, although there are substantial distinctions in their taxonomy.

The climate was less warm and more humid at the Gornaya flora time as compared to the Barykov time. Angiosperms predominant in the Gornaya flora include genera *Peculnea*, *Trochodendroides*, *Celastrinites*, *Renea*, *Dyrana*, *Platanus*, *Viburnum*, and *Corylus*. Conifers are subordinate as compared to angiosperms and mainly represented by the typical Late Cretaceous forms (*Sequoia minuta*, *Cryptomerites*, *Elatocladus smittiana*, *Parataxodium*) and by some younger elements (*Metasequoia* and *Glyptostrobus*). *Ginkgo* and cycadophytes (*Nilssonia*, *Encephalartropsis*), which probably formed monodominant associations, become common in the flora. Ferns are represented by scarce remains. The amount of large-leaved Platanaceae increases in the Gornaya flora as compared to the Barykov flora, whereas the abundance of species with entire leathery leaves decreases, thus pointing to a certain humudification and cooling of the climate.

The Rarytkin flora (the late Maastrichtian–Danian) indicates a further cooling and humidification of the climate. The Rarytkin species had larger leaves as compared to the Gornaya flora, and many of them were covered with serrate leaves. Large-leaved Platanaceae are usual components of the flora. Conifers dominate over flowering plants in the abundance and are represented in burials by branch-deciduous forms. Cycadophytes disappear from the flora, while ferns become more frequent.

*Corylus beringiana*, some species of the *Trochodendroides* genus, *Metasequoia*, *Glyptostrobus*, and *Mesocyparis* dominate in the Rarytkin flora. *Taxodium*, *Platanus raynoldsii*, *Quercus groenlandica*, *Arthollia rarytkensis*, *Rarytkinia*, *Celastrinites*, *Nyssa*, *Platimelis*, *Viburnum*, *Vitis*, *Quereuxia*, and *Haemanthophyllum* are representative plants in the flora. *Gingko* remains are very rare. Ferns are represented by genera *Onoclea* and *Coniopteris*.

In total, six floral stages are distinguished in the Upper Cretaceous deposits of northeastern Russia. In all instances, transformations of flora resulted from changes in climatic conditions. Each event can be considered as an ecological crisis. The reorganization of flora proceeds rapidly and usually leaves no traces in the geologic record. Most elements of the preceding flora (mainly stenobiontic dominant taxons of climax communities) become extinct parallel to the species migration and diversification (Krasilov, 1985; Nesov and Golovneva, 1987). The subsequent coevolution results in the appearance of new plant communities adapted to new environments that have already existed in a stable state for millions of years. Life spans of the Late Cretaceous floras can be as long as 2–4 million years (Table 3).

The nature of changes in the flora at the stage boundaries depends on the nature of the climatic change, its trend and sharpness. Two principal types of changes in the flora can be distinguished: (1) changes occurring under the deterioration of climatic conditions; (2) changes occurring under conditions of optimal relation between the humidity and temperature regime vital for plants.

Climatic changes of the first type are responsible for the decline of the preceding flora, sometimes at a high taxonomic level, as well as for a decrease in the species diversity and an increase in the abundance of dominant species in the successive flora. Conifers often become important components of vegetation.

With the onset of the optimal climatic conditions, the extinction is less dramatic, the species diversity increases, and many new taxons appear in a new flora. Dominant elements of the preceding flora do not die out under optimal conditions but gradually loose their importance, whereas the deterioration of climatic conditions results in a drastic extinction of the dominant species (stenobiontic to a greater or lesser degree). In the latter case, assectators or pioneer species (i.e., various eurybiontic organisms) of the preceding stage or newly formed taxons become dominant in the new flora (Fig. 3).

Characteristic examples of the first type are changes in the flora at the end of the Maastrichtian (the Gornaya–Rarytkin assemblages) in response to the climatic cooling, as well as changes at the boundary between the Coniacian and Santonian (the Kaivayam–Valizhgen

assemblages), which were caused by aridization. Changes of the second type are more evident in the transition from the Penzhina to the Kaivayam assemblage, when a humid and cool climate gave way to similarly humid but warmer conditions.

The Grebenka flora, which is characterized by maximum species diversity, is most optimal among other Late Cretaceous floras. The warming of the climate stimulated the intense formation of new taxons. The first thriving phase of older flowering plants and their prevalence in the flora falls within the Grebenka time.

Major reorganizations in the Late Cretaceous floras related to drastic extinction of taxons took place under deteriorating climatic conditions at the boundary between the early and late Turonian, between the Coniacian and Santonian, and again between the middle and late Maastrichtian.

Along with common features, each reorganization has its own peculiarities. Both the Turonian and Maastrichtian events occurred as a result of cooling. A considerable extinction preceded the Penzhina and Rarytkin floras. In addition to extinction, a southward migration of thermophilic taxons took place in the Penzhina time. In the late Maastrichtian–Danian, the cooling was more intense and spread over a vast territory extending to southern Mongolia. Moreover, the latitudinal thermal gradient and seasonal fluctuations increased at that time (Wolfe and Upchurch, 1987). The preceding middle Maastrichtian Gornaya flora, which evolved under stable semi-arid conditions, was characterized by very high specialization of taxons. As a consequence, transformations in floras were more significant than in the Turonian time. The replacement of flowering plants went parallel to reorganization of gymnosperms. A representative group of conifers that became extinct at that time included *Parataxodium*, "*Cephalotaxopsis*," the Late Cretaceous species of *Sequoia*, *Elatocladus smittiana*, and *Araucarites*, which dominated in floras almost throughout the Late Cretaceous and safely survived the Turonian crisis.

A feature common to both reorganizations is the succession of many eurytopic species with wide geographic and stratigraphic ranges from the preceding floras. Newly-formed taxons of the Penzhina and Rarytkin floras were also characterized by wide geographic ranges and poor ecologic specialization. As a consequence, floras of the cold and humid climatic epochs become less differentiated in the geographic aspect and more homogeneous in composition. These periods in the flora evolution are especially favorable for phytostratigraphic correlations over vast territories.

During the flora transformation in response to the deterioration of climatic conditions, accessory taxons, which were already present in the preceding flora, become dominant. These groups underwent substantial diversification under new conditions. Platanaceae and *Trochodendroides* are of this sort in the Penzhina flora, like Betulaceae and *Trochodendroides* in the Rarytkin

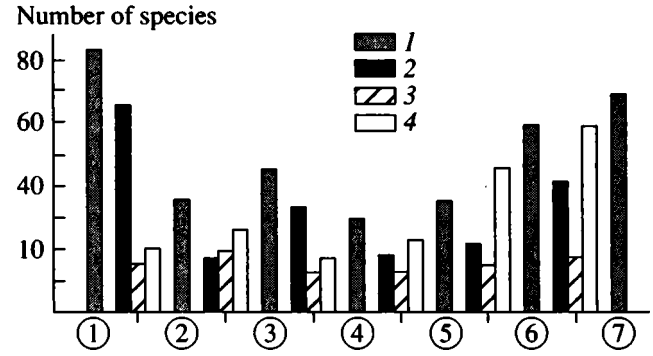


**Fig. 4.** Changes in the species diversity of the Cretaceous floras from northeastern Russia: (*1*) total amount of species in the flora; (2) extinct species; (*3*) transitory species; (*4*) reappearing species. Numerals in circles indicate fossil floras: 1, Grebenka; 2, Penzhina; 3, Kaivayam; 4, Valizhgen; 5, Barykov and Verkhnyaya Bystraya; 6, Gornaya and Kakanaut; 7, Rarytkin and Koryak.

flora. However, the Penzhina flora has no species which originate from the Grebenka flora unchanged to become dominant. In this case, dominant plants represent new species of older genera, or even new genera. Transient eurybiontic and widespread species play a subordinate part in the Penzhina flora. Dominant species in the Rarytkin flora are those which have already existed in the Gornaya flora time. The appearance of new taxons in the Rarytkin time is considerable, but most of them are poorly abundant.

Another peculiarity of the Gornaya–Rarytkin flora reorganization is the resulting increase in species diversity despite the predominance of eurytopic species with wide geographic ranges, and structural simplification of the plant assemblage displaying the prevalence of 4–5 species. The reason is that after the significant Maastrichtian extinction, which touched all groups of plants, and after the persisting abiotic stress, the Rarytkin ecosystems retained a pioneer structure with numerous free ecological niches. This factor appeared to be favorable for a new taxonomic radiation (Golovneva, 1994). As a result, 75% of the species are new in the late Maastrichtian–Danian Rarytkin flora, which exceeds the amount of extinct taxons, representing 62% of the Gornaya flora (Table 5, Fig. 4).

Changes in the flora composition at the boundary between the Coniacian and Santonian (transition from the Kaivayam to Valizhgen flora), which were concurrent with the climate aridization, are also considerable but somewhat different in nature. Both cold-resistant large-leaved Platanaceae and hydrophilous entire-leaved thermophilic forms died out during this event. This reduced the diversity of species, especially of flowering plants, a share of which decreases from 60 to 37% (this is close to their amount in the Cenomanian). The simultaneous simplification of the assemblage structure turned it from polydominant to monodominant. Only three species (*Macclintockia crenata*,

*Sequoia* ex gr. *reichenbachii*, and *Nilssonia*) prevail in the sites of the Valizhgen flora. Their assemblage is of the monodominant type. Other species are rare. The role of the Santonian migrations was insignificant. Taxons that disappeared from the assemblage are younger conifers (*Metasequoia*, *Thuja*) and some flowering plants, which found refuge in more humid mountain regions of the Okhotsk–Chukchi volcanogenic belt. Genetic relations between the Valizhgen and Kaivayam assemblages are weak. Dominant elements of the Valizhgen flora represent new taxons. A warm semi-arid steady climate was favorable for specialization of taxons. This increased the geographic differentiation of the Boreal–Pacific floras during the Santonian, Campanian, and middle Maastrichtian. The endemic nature of floras becomes more pronounced in the middle Maastrichtian (Gornaya time), when the floras of adjacent provinces had no identical genera among angiosperms (Golovneva, 1994). A high specialization of the dominant Valizhgen taxons was responsible for extinction of most species of this kind at the end of the stage despite the humidification of the Barykov time.

Thus, climatic conditions and the character of their changes affect directly not only the species diversity and ecosystem structures, but also the degree of geographic differentiation of floras and the extinction and generation of taxons, when evolutionary stages of floras change one another. Epochs with a warm and humid climate are characterized by a high species diversity, the prevalence of polydominant assemblages of highly specialized species, and an abundance of new taxons and moderate geographic differentiation of floras. The Grebenka and, to a lesser degree, the Kaivayam assemblages exemplify floras of this kind for the Late Cretaceous of northeastern Russia.

Epochs with a warm and steadily arid climate (the Santonian–middle Maastrichtian) are characterized by some decrease in the species diversity and an increase in the extent of prevalence of some species in plant assemblages. They are also responsible for a high specialization of taxons and the highly endemic nature of floras in individual regions. Many stenobiontic species become extinct at the end of such epochs, even under the effect of insignificant climatic changes. The inheritance degree in their evolution is low.

In epochs with a cold and humid climate with pronounced seasonal fluctuations, the species diversity also decreases and the extent of prevalence of some species increases. In such a case, eurytopic species with wide geographic and stratigraphic ranges (the Penzhina and Rarytkin regional stages) are dominant. Due to the lower specialization, the taxons survive the ecological crises better at the boundary between floral epochs, and the degree of their extinction at the end of the crises is usually less impressive.

It should be pointed out that the flora as a whole and individual plant assemblages are more sensitive to the humidity variations and, to a lesser extent, to the temperature changes.

The geographic isolation of paleofloras is a less effective factor of differentiation than the climatic changes. In particular, the late Turonian was a time of the major Cretaceous transgression that caused the maximum isolation of different regions (Naidin *et al.*, 1986). At the same time, the phytogeographic differentiation was remarkably low, just a little higher than in the late Maastrichtian–Danian.

The analysis of the extinction and speciation in various floral assemblages of the Late Cretaceous (Table 5, Fig. 4) shows that in all of them the amount of newly-formed species at the beginning of the stage is approximately equal to the amount of species that become extinct at the end. This is true of the climatic pessimums and optimums when both the appearance of new species and extinction are significant. It testifies that speciation is a principal reaction of flora to climatic changes. Diversification and extinction at the generic or higher taxonomic levels depend to a lesser degree on the environmental factors.

In general, the extent of flora renewal at every stage is high. Forty five to eighty percent of species become extinct at the end of a stage. In this respect, the Cretaceous–Paleogene boundary event (consistent with the Gornaya–Rarytkin flora reorganization in northeastern Russia when 62% of species became extinct) is not unique among other Cretaceous crises in the flora evolution.

When analyzing the extinction at the boundaries between floral stages, a surprisingly steady state of the transient flora drew our attention. Amounts of transitory species during various crises are much the same regardless of the general diversity of floras and the character of their alternation. A transitory group usually comprises the eurybiontic taxons and para-aquatic plants less dependent on climatic conditions. However, in the Late Cretaceous floras of northeastern Russia, conifers constitute the major proportion of transitory species, which are dominant or subdominant forms in all floras. Among them, there are some *Sequoia* species, "*Cephalotaxopsis*" *heterophylla*, *Elatocladus smittiana*, *Pityophyllum* sp., *Parataxodium*, and *Cupressinocladus cretaceus*, as well as *Gingko adiantoides* and *Nilssonia serotina* in the thermophilic flora. This may indicate a considerable stability of these gymnosperm assemblages, which persisted throughout the Late Cretaceous and probably dominated in the valley, slope, and swamp vegetation. However, proportions of species in various floras of the Late Cretaceous conifers usually change depending on the climate, or a modest speciation occurs among the already existing genera. *Sequoia* ex gr. *reichenbachii* appears in the Santonian, while *S. factigiata* becomes representative in the Campanian, although it is typical of the southern floras of the Far East throughout the Cretaceous. The additional cuticular analysis is likely be able to show that the spe-

cies composition of the Late Cretaceous conifers was more diverse. Nevertheless, these plants did not change in their outward appearance. In the late Maastrichtian time, the Late Cretaceous assemblages of conifers were replaced by recent genera *Metasequoia*, *Glyptostrobus*, and *Taxodium*, whose assemblages were widespread throughout the Tertiary and still survive, though in a slightly changed form.

The evolution of angiosperms, unlike conifers, was very intense. About 60–90% of flowering plant species were replaced stage by stage, and the proportion of the replaced genera and families is significant as well. The abundance of flowering plants gradually increased during the Late Cretaceous up to 70% by the end of the period. The diversity and abundance of ferns concurrently decreased (from 23% in the Grebenka flora to 6% in the Rarytkin flora). Such a decrease cannot be explained by climatic changes and is likely related to the replacement of ferns by flowering plants and probably by some conifers from their former habitat areas populated by nonforest plant groups. Non-forest ferns and cycadophytes vanished by the end of the Cretaceous in northeastern Russia and in many other regions (Krassilov, 1995).

A share of relict ginkgoaleans, cycadophytes, and ferns in the Late Cretaceous floras gradually decreased. *Podozamites*, czekanowskialeans, *Baiera*, and *Sphenobaiera* disappeared at the end of the Grebenka, while caytonialeans (*Sagenopteris*) vanished at the end of the Valizhgen. The Gornaya flora was the last to incorporate a significant part of typical Late Cretaceous conifers and cycadophytes. At the same time, *Ginkgo* forms lost their dominant role in assemblages and became accessory in later floras. As far as angiosperms are concerned, their taxonomic renewal at the boundary between the middle and late Maastrichtian was not greater than in the Late Cretaceous. The amount of recent angiosperm genera increases from a third to a half of the assemblage composition and some become dominant, but the importance of extinct genera still remains high. More important taxonomic reorganizations among flowering plants occurred in the Late Paleocene–Early Eocene and were related to the progressive warming of the climate and to evolution of Ulmaceae, Betulaceae, Fagaceae, Juglandaceae, Hamamelidaceae, and other families, which are the principal constituents of the current deciduous floras of the northern hemisphere.

## ACKNOWLEDGMENTS

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# The Bed-by-Bed Standard of the Silurian for Eastern Siberia

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**Abstract**—The bed-by-bed standard of the Silurian for Eastern Siberia is elaborated for the first time in order to correlate all natural and borehole sections of the region with the detailed standard sections in other type areas. The standard includes stratotypes of regional horizons, subhorizons, chronozones, and beds distinguished in the Silurian stratigraphic sequences of the Moiero, Turukhansk, and Norilsk stratotype areas of Eastern Siberia. All the beds and interbeds graphically shown in the composite standard section are supplemented with data on distribution ranges of different faunal groups. Their lithologic and taphonomic characteristics are also presented.

**Key words:** *Silurian, standard, Siberia, horizon, chronozone, bed*

The bed-by-bed standard of the Silurian for Eastern Siberia is based on two sections exposed in the Moiero River basin (Llandoverian and Wenlockian series) and along the Kureika River valley (Gorstian Stage), and also on the Ludfordian–Pridolian section recovered by Borehole TT-1 at the eastern coast of Pyasino Lake (Figs. 1 and 2). Systematic investigation of the standard was initiated in 1972, when the working group of specialists in sedimentology, ecology, lithology, and paleontology was established to study the Silurian sedimentation basin of Eastern Siberia in all geological aspects.

Summarizing the early results, this scientific team issued a book entitled "The Silurian of Siberian Platform: New Regional and Local Stratigraphic Subdivisions" (Tesakov *et al.*, 1979). At that time, the Silurian of the region was divided into six horizons (Moiero, Khaastyr, Agidy, Khakoma, Tukal, and Postnichnyi), and the Mukdeken Horizon overlain by the Zub Horizon was recognized as the basal one in the Devonian. Type sections of these units were described in detail a little later (Tesakov *et al.*, 1980, 1985; Predtechenskii and Tesakov, 1982).

At present, nearly all outcrops and numerous borehole sections of Silurian rocks are successfully studied in Eastern Siberia (the Siberian platform and Taimyr). As a result of their analysis, the former regional stratigraphic scheme (Tesakov *et al.*, 1979, 1983) was revised to a certain extent and approved as valid for all of Eastern Siberia. In addition, it was detailed up to the subhorizon and chronozone levels (Tesakov *et al.*, 1992, 1996a, 1996b, 1998). The established stratotypes of regional chronozones are used as units constituting stratotypes of subhorizons, horizons, and superhorizons in the stratigraphic ranges of the latter authorized after the revision.

In the latest scheme, the superhorizons are divided into horizons consisting of two (lower and upper), or three (lower, middle, and upper) subhorizons. The latter include several (two to eight) regional chronozones labeled with indices 1, 2, 3, etc. In the general succession, there are 54 sequentially numbered regional chronozones usable as correlation levels for geological mapping and stratigraphic calibration of geological sections and profiles. The ranges of the regional chronozones also outline the limits of the respective biointervals.

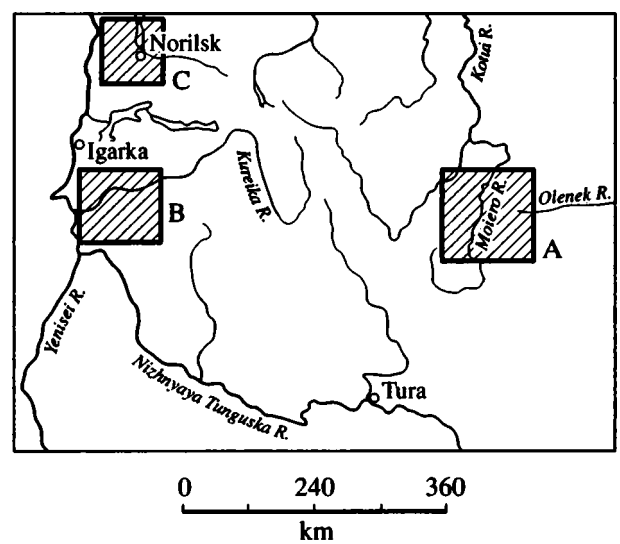


**Fig. 1.** Type areas studied to elaborate the bed-by-bed standard of the Silurian for Eastern Siberia: (A) Moiero region; (B) Turukhansk region; (C) Norilsk region.

Lithologic and taphonomic description of the bed-by-bed Silurian standard for Eastern Siberia (Fig. 3) is presented below. The chronozones are indicated first by their ordinal number in the standard under consideration followed by the numerical index designating its position in a given subhorizon. The stratotype bed numbers for chronozones and their thickness are also indicated.

*Moierokan Horizon*, mr, (110.0 m).

**Lower Moierokan Subhorizon** (lower and basal middle Rhuddanian), $mr_1$, (54 m).

(1) *Chronozone 1*, $mr_{1(1)}$: Bed 82/1 (0.5 m).

Bed 82/1: dark-gray bituminous medium-platy limestones with a minor clay admixture, pyrite geodes, and black shale interlayers. The bed encloses rare dispersed and clustered organic remains, including the dominant *Alispira gracilis* (0.5 m).

(2) *Chronozone 1*, $mr_{1(2)}$: Beds 82/2 and 82/3 (1.8 m).

Bed 82/2: black bituminous thick-platy limestones with a minor admixture of clay and pyrite geodes. Abundant organic remains of the bed occur as stratal accumulations. *Geisonoceras kureikense* is the dominant form (0.2 m).

Bed 82/3: basal black shales with the fine horizontal lamination and large pyrite-marcasite nodules grade upward into the clayey limestones also displaying the horizontal lamination. Accumulations of rare organic remains are confined to bedding planes. *Glyptograptus tamariscus* is dominant (1.6 m).

(3) *Chronozone 3*, $mr_{1(3)}$: Beds 82/4–82/7 (5.0 m).

Beds 82/4 and 82/5: frequent (1–4 mm) intercalation of light-gray marls and gray calcareous shales, packets of which are 10–20 cm thick and separated by interlayers (2–3 cm) of gray dolomitic limestones. Rare and small organic remains are dispersed along the bedding planes. The dominant form is *Metabolograptus moyeroensis* (1.1 m).

Beds 82/6 and 82/7: gray clayey-dolomitic limestones (2–4 cm thick) showing the wavy lamination and regularly alternating with gray clayey marls (2–3 cm thick). Small and frequent organic remains mostly characteristic of marls are buried along the bedding planes. Dominant forms are *Metabolograptus moyeroensis* in the Bed 82/6 and *Paraclimacograptus innotatus* in the Bed 82/7 (3.4 m).

(4) *Chronozone 4*, $mr_{1(4)}$: Bed 82/8 (5.0 m).

Bed 82/8: gray clayey-dolomitic limestones (2–3 cm thick) displaying the gentle wavy lamination and intercalated with gray calcareous shales (1–2 cm thick). Rare and small organic remains are dispersed over the bedding planes. *Coronograptus cyphus angustus* is dominant among fossils (5.0 m).

(5) *Chronozone 5*, $mr_{1(5)}$: Beds 82/9 and 82/10 (8.5 m).

Beds 82/9 and 82/10: frequent (1–3 cm) alternation of limestones and shales like that in the Bed 82/8. Rare

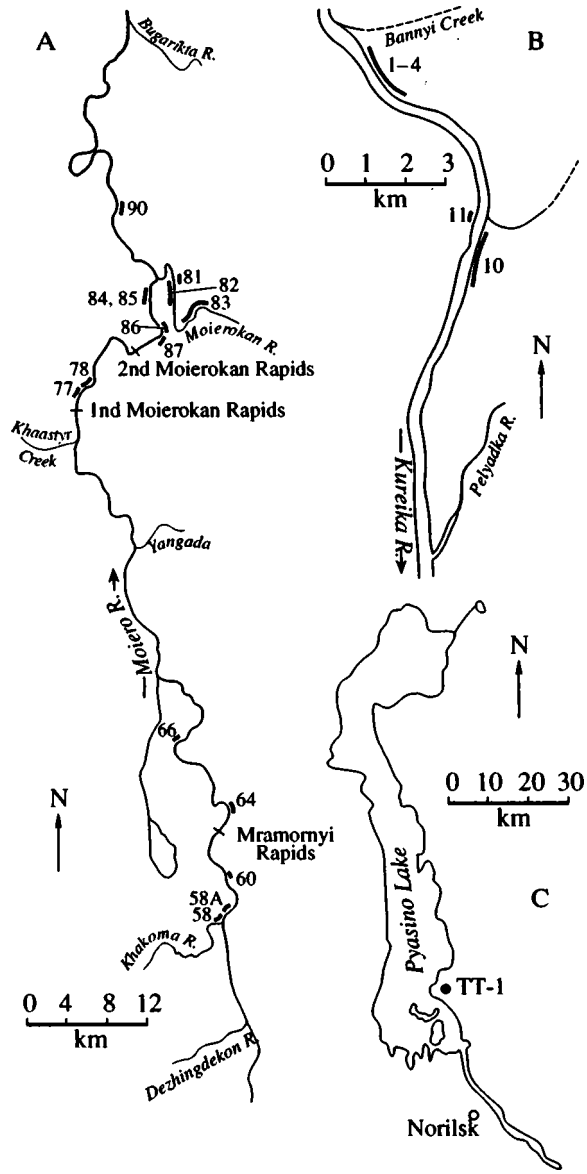


**Fig. 2.** Type section localities studied to elaborate the bed-by bed standard of the Silurian for Eastern Siberia. (A) Moiero district, sections of the Moierokan River: (82) left-hand bank 3.0 km upstream of the mouth, (83) right-hand bank 5.0 km upstream of the mouth; sections of the Moiero River: (78) left-hand bank 4.0 km downstream of the Khaastyr Creek mouth, (90) right-hand bank 5.0 km downstream of the Moierokan River mouth, (77) left-hand bank 2.0 km downstream of the Khaastyr Creek mouth, (84, 85) left-hand bank 2.0 km upstream of the Moierokan River mouth, (86) right-hand bank 2.7 km upstream of the Moierokan River mouth, (87) right-hand bank 4.5 km upstream of the Moierokan River mouth, (60) left-hand bank 3.0 km downstream of the Khakoma River mouth, (58A) left-hand bank 0.7 km downstream of the Khakoma River mouth, (58) left-hand bank near the Khakoma River mouth, (64) right-hand bank 3.0 km downstream of the Mramornyi Rapids, (66) left-hand bank 9.0 km downstream of the Mramornyi Rapids. (B) Turukhansk region, sections of the Kureika River: (11) Nizhnie Shcheki locality, right-hand bank, (1–4) Verkhnie Shcheki locality, left-hand bank, (10) Nizhnie Shcheki locality, left-hand bank. (C) Norilsk region, eastern coast of the Pyasino Lake, borehole TT-1.

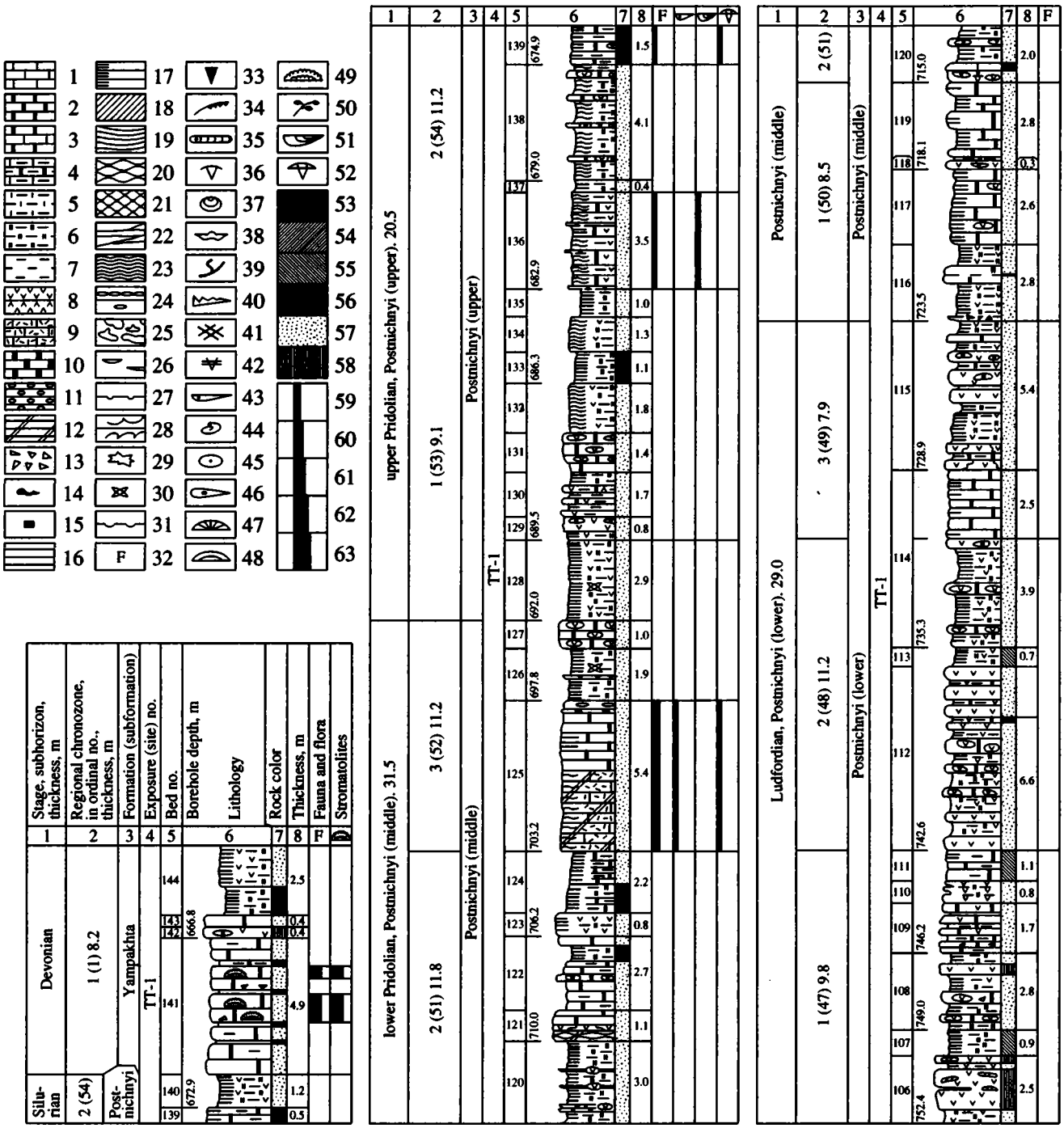


**Fig. 3.** The bed-by-bed standard of the Silurian for Eastern Siberia: (1) limestone; (2) dolomite; (3) dolomitic limestone or limy dolomite; (4) clayey limestone and dolomite; (5) marl; (6) domerite; (7) shale; (8) gypsum and anhydrite; (9) detrital limestone; (10) bituminous limestone; (11) ooliths; (12) secondary dolomitization and calcification; (13) authigenic breccia; (14) cherty nodules; (15) pyrite; (16) coarse and medium horizontal bedding; (17) fine horizontal bedding and lamination; (18) cross lamination; (19) wavy lamination; (20) coarse-grumous structure; (21) fine-grumous structure; (22) grumous-laminated structure; (23) tubercular bedding; (24) nodules and nodular interlayers; (25) nodules of irregular shape; (26) lenses; (27) exsiccation cracks; (28) ripple marks; (29) cavities; (30) hieroglyphs; (31) diastems; (32) fauna; (33) hyolithids; (34) graptolites; (35) cephalopods; (36) trilobites; (37) inarticulate brachiopods; (38) articulate brachiopods; (39) mudeaters; (40) conodonts; (41) bryozoans; (42) tentaculitids; (43) ostracodes; (44) gastropods; (45) crinoids; (46) rugoses; (47) tabulates; (48) stromatoporoids; (49) stromatolites; (50) algae; (51) pelecypods; (52) eurypterids; (53) black and dark-gray rocks; (54) red, brown, or lilac rocks; (55) green rocks; (56) variegated rocks; (57) gray rocks; (58) light-colored rocks; (59) single fossils; (60) rare fossils; (61) frequent fossils; (62) widespread fossils; (63) abundant fossils.

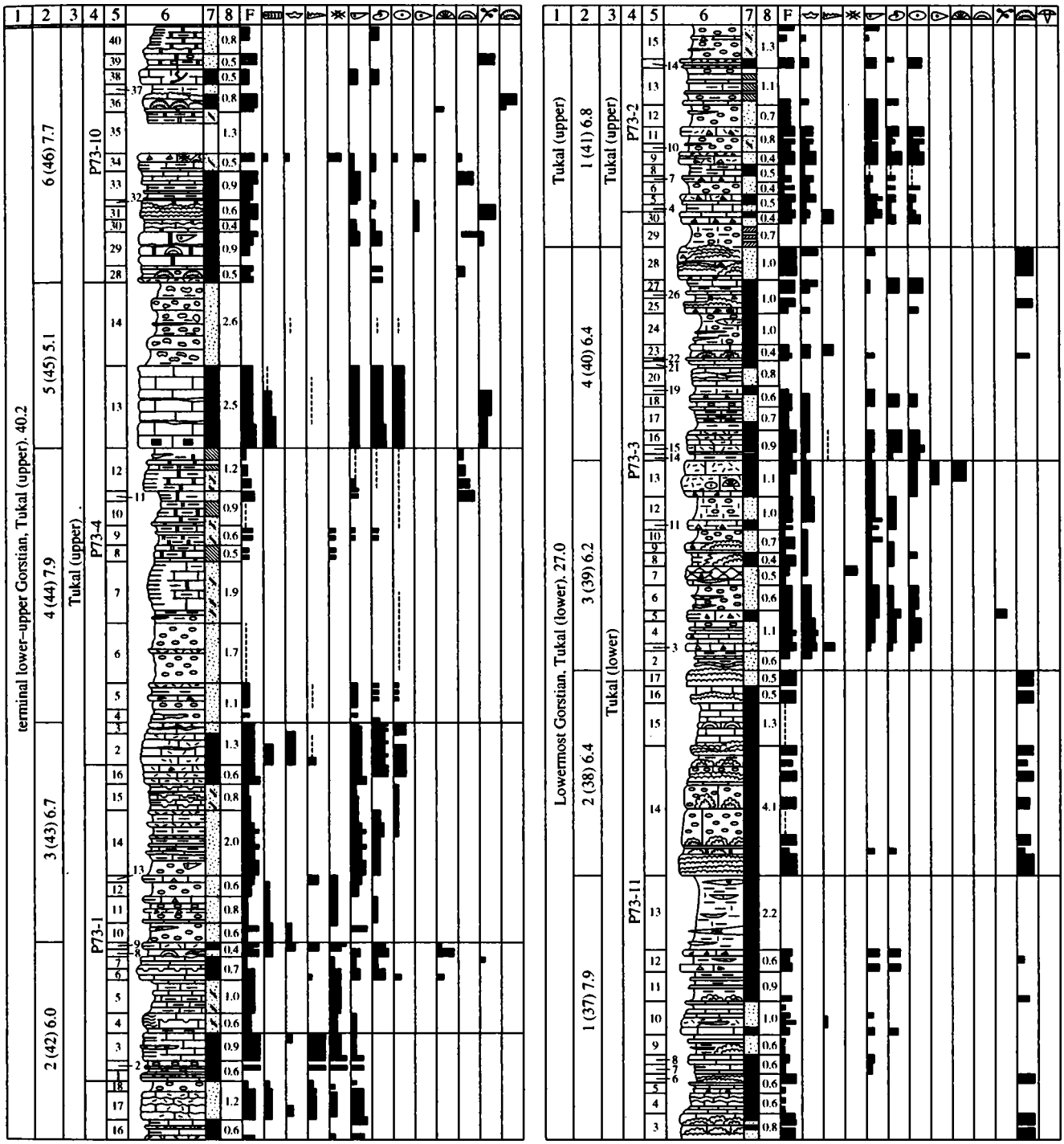


**Fig. 3.** (Contd.)

and small organic remains are scattered over the bedding planes in shales. *Coronograptus cyphus angustus* again is dominant (8.5 m)

(6) *Chronozone 6*, $mr_{1(6)}$: Beds 82/11 and 82/12 (8.3 m).

Beds 82/11 and 82/12: gray clayey-dolomitic limestones of the unit are medium-platy (up to 6 cm), show small tubercles at their bedding planes, and are separated by thin interlayers (1–3 cm) of greenish gray to black marls. Basal portions of limestone interlayers are burrowed by mudeaters (burrows up to 1 cm in diameter), and *Paraclimacograptus innotatus* is dominant among small and dispersed fossil remains (8.3 m).

(7) *Chronozone 7*, $mr_{1(7)}$: Bed 82/13 (11.0 m).

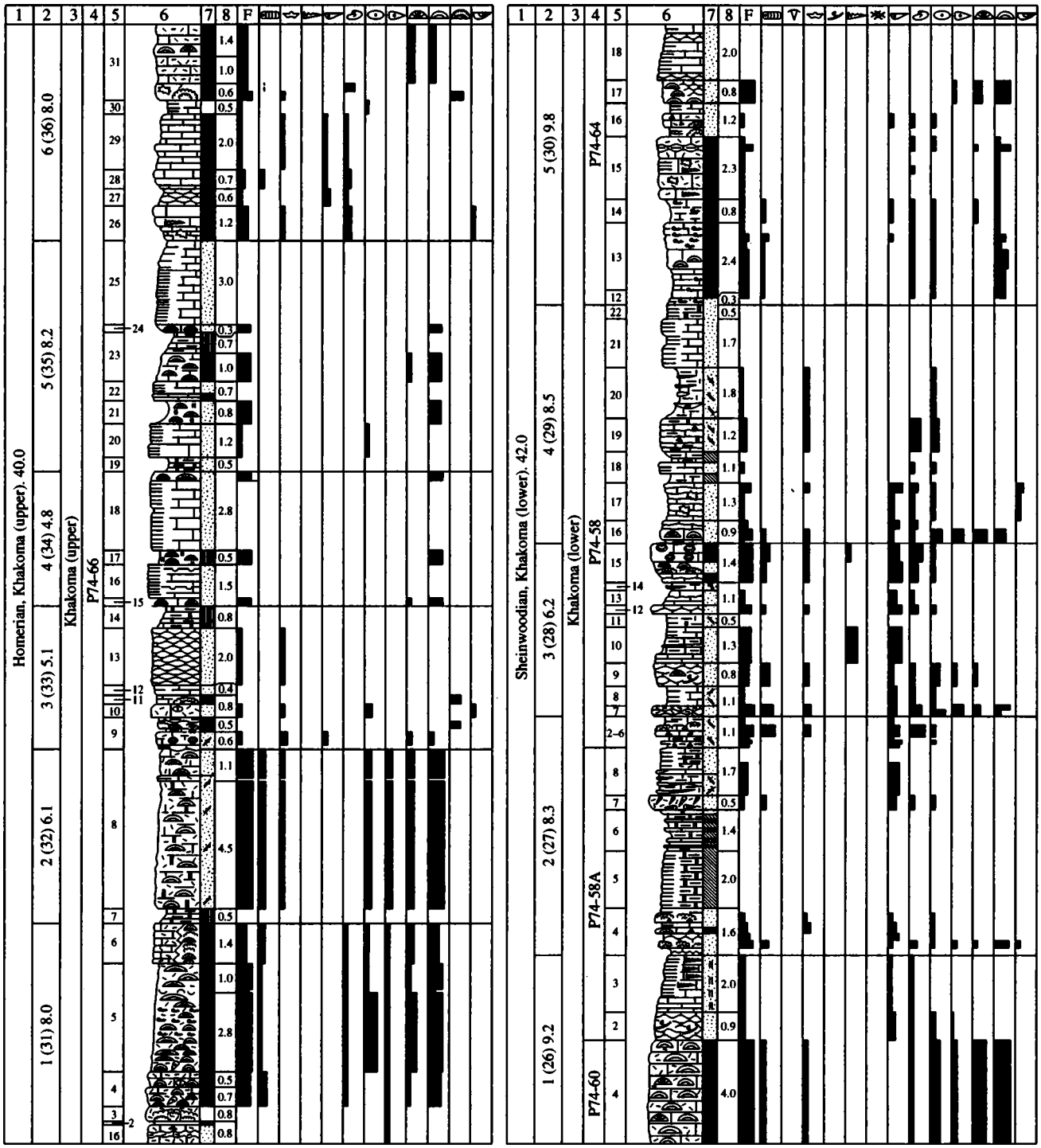


**Fig. 3.** (Contd.)

Bed 82/13: gray dolomitic tubercular-platy (plates 2–3 cm large) limestones with rare and thin shale laminae. Frequent small fossil remains in the rocks are mainly represented by *Acernaspis superciliexcelsis* (11.0 m).

(8) *Chronozone 8*, $mr_{1(8)}$: Beds 82/14–82/16 (13.7 m).

Beds 82/14 and 82/15: layers of gray clayey-dolomitic tubercular-platy (plates 1–2 cm thick) limestones alternating with gray clayey-dolomitic platy (plates 2–3 cm thick) limestones, displaying the fine horizontal lamination, and with gray marl interlayers up to 1.0 cm thick. Organic remains are small and frequent, predominantly represented by *Acernaspis superciliexcelsis* (4.9 m).

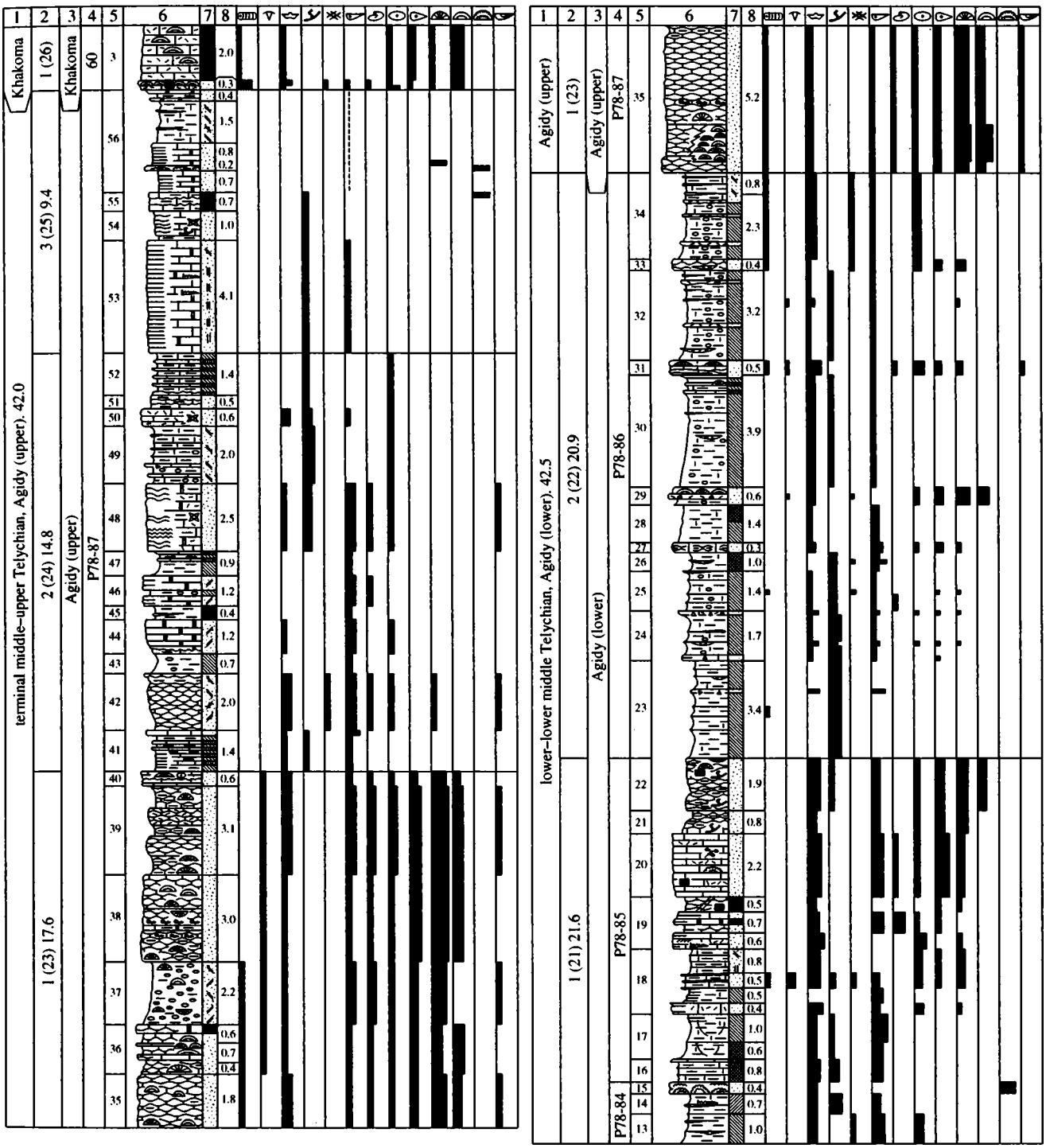


**Fig. 3.** (Contd.)

Bed 82/16: gray clayey-dolomitic thin-platy (1.5–2.0 cm) limestones with tubercular lamination and interlayers of greenish gray marls enclose isolated interbeds (1.5 cm thick) of greenish gray grumous limestones. *Clorinda undata* is the dominant form of fossils abundant in the bed (8.8 m).

**Upper Moierokan Subhorizon** (upper middle Rhuddanian–lower Aeronian), $mr_2$, (5.70 m).

(9) *Chronozone 1*, $mr_{2(1)}$: Beds 82/17 and 82/18 (11.8 m).

Beds 82/17 and 82/18: packets of gray grumous clayey-dolomitic limestones enclose rare interlayers

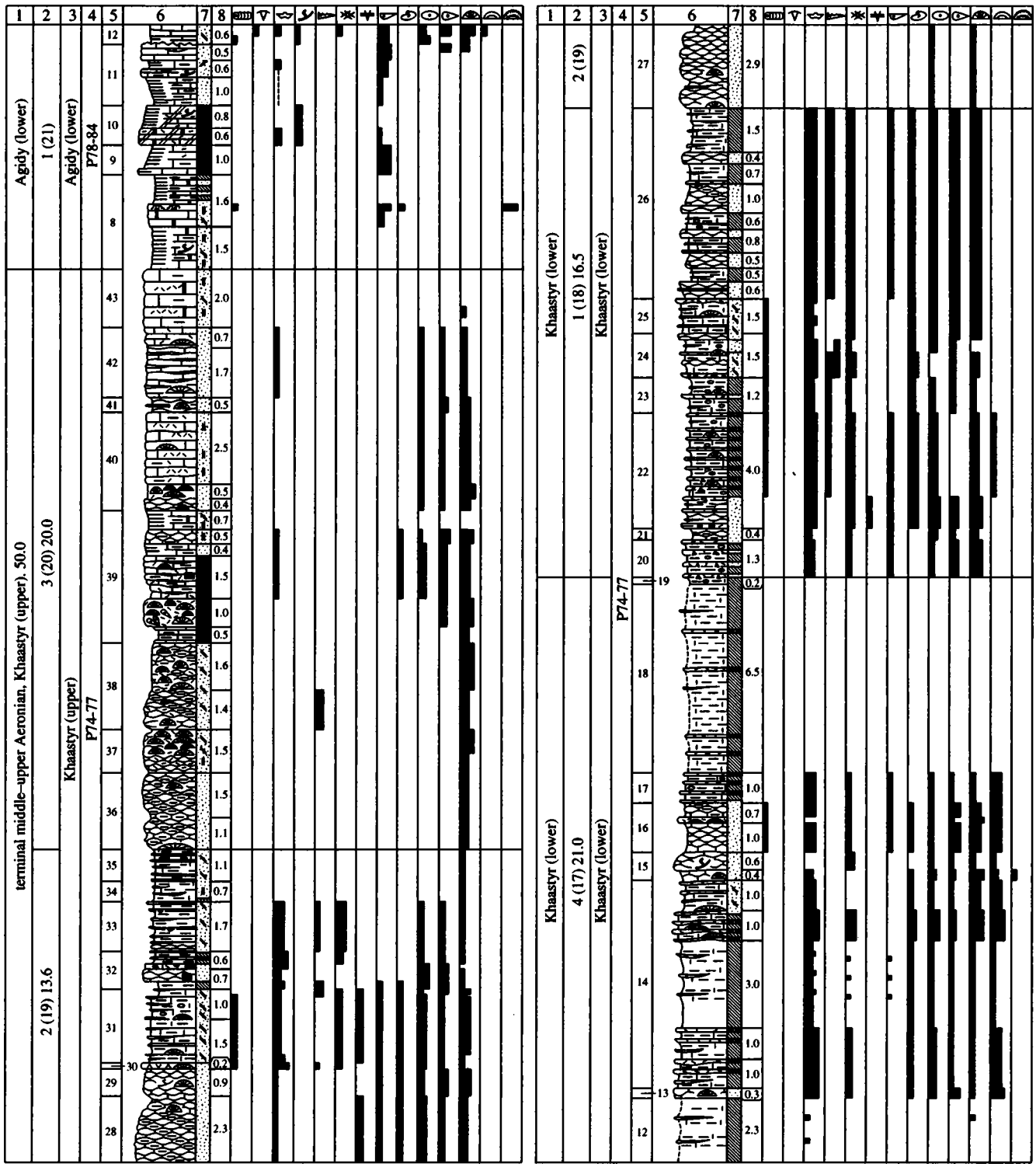


**Fig. 3.** (Contd.)

(1–3 cm thick) of greenish gray detrital limestones and are intercalated with packets (0.3–0.5 m thick) of gray clayey limestones displaying the tubercular lamination and loose laminated marl interlayers. The dispersed or clustered organic remains of the unit include *Sibiritia wiluiensis* as the dominant species (11.8 m).

(10) *Chronozone 2*, $mr_{2(2)}$: Beds 82/19 and 82/20 (10.5 m).

Beds 82/19 and 82/20: gray to dark-gray grumous clayey-dolomitic limestones rich in detrital material. Organic remains small to medium in size are either dis-

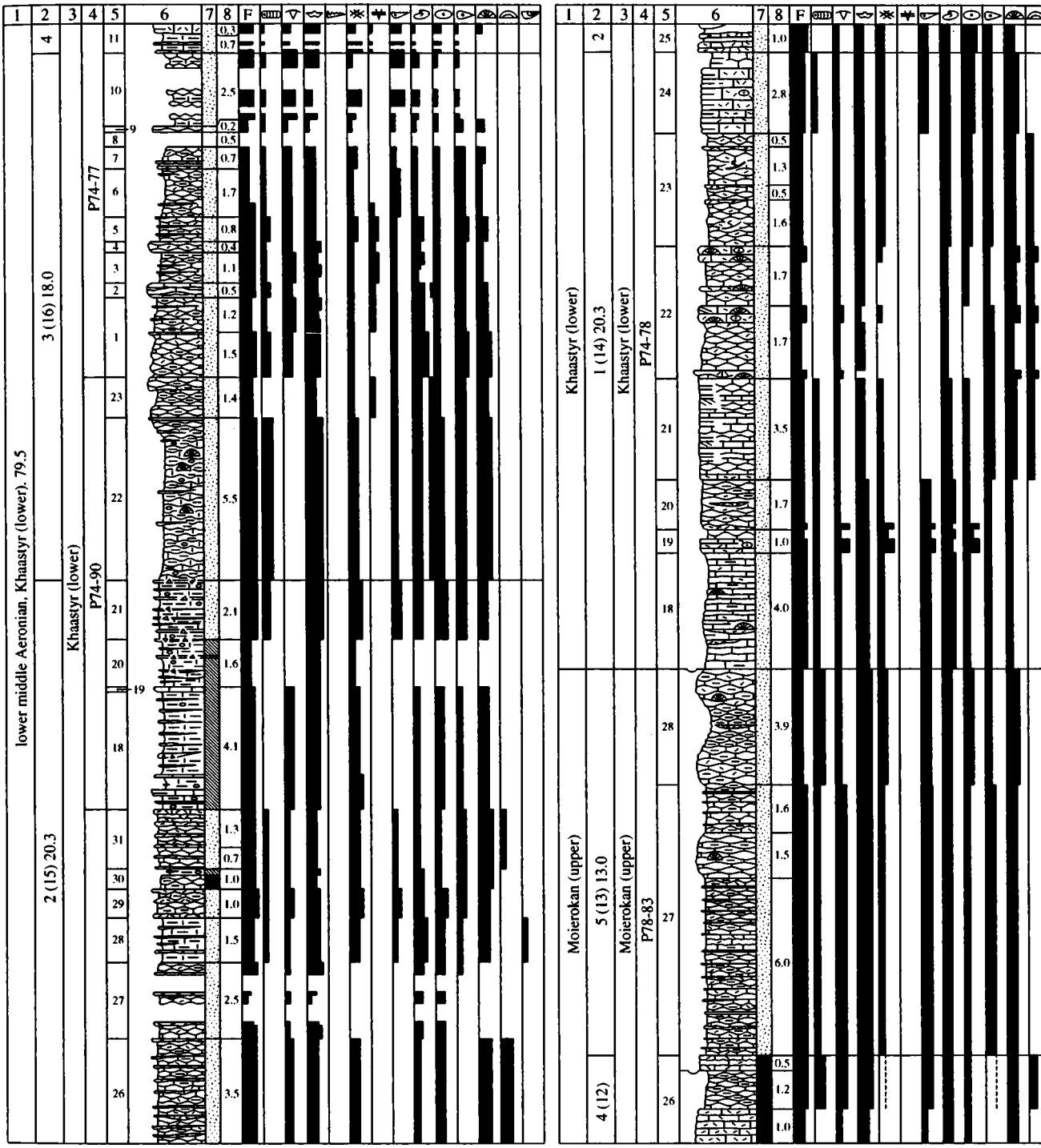


**Fig. 3.** (Contd.)

persed, or occur in accumulations. Their dominant representative is *Septatrypa antiquata*. Bioturbations of the zoophycos type are also recorded (10.5 m).

(11) *Chronozone 3*, $mr_{2(3)}$: Beds 82/21–82/23 (10.2 m).

Beds 82/21–82/23: gray grumous clayey-dolomitic limestones with abundant detritus and lenticular laminae of greenish gray clayey marls. The rocks show bioturbation of the zoophycos type and mudeater tracks. Small- and medium-sized organic remains occur in the dispersed or clustered state. Dominant forms are *Sibiritia wiluiensis* in the Bed 82/21, *Calamopora alveolaris* in the Bed 82/22, and *Zygospiraella duboisi* in the Bed 82/23 (10.2 m).

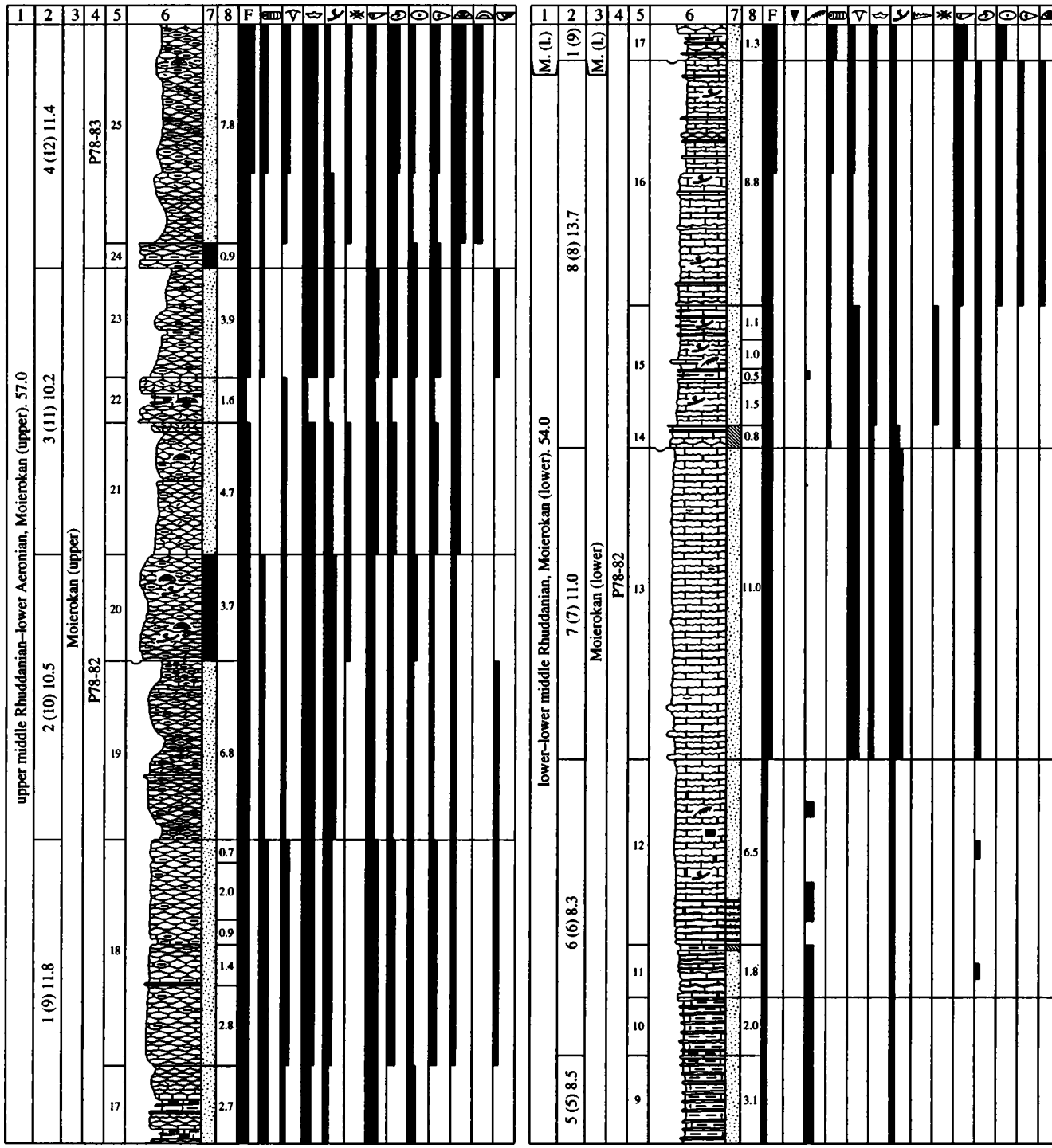


**Fig. 3.** (Contd.)

(12) *Chronozone 4*, $mr_{2(4)}$: Beds 83/24–83/26 (11.4 m).

Beds 83/24–83/26: grumous clayey-dolomitic limestones gray at the base (Bed 83/24) and dark-gray at the top (Bed 83/26), are rich in detrital material and enclose frequent laminae of marls. Abundant, though dispersed fossils medium in size include such dominant species as *Isorthis neocrassa* (Bed 83/24), *Calamapora alveolaris* (Bed 83/25), and *Septatrypa antiquata* (Bed 83/26). Bioturbation of the zoophicos type occurs (Bed 83/25). 11.4 m.

(13) *Chronozone 5*, $mr_{2(5)}$: Beds 83/27 and 83/28 (13.0 m).

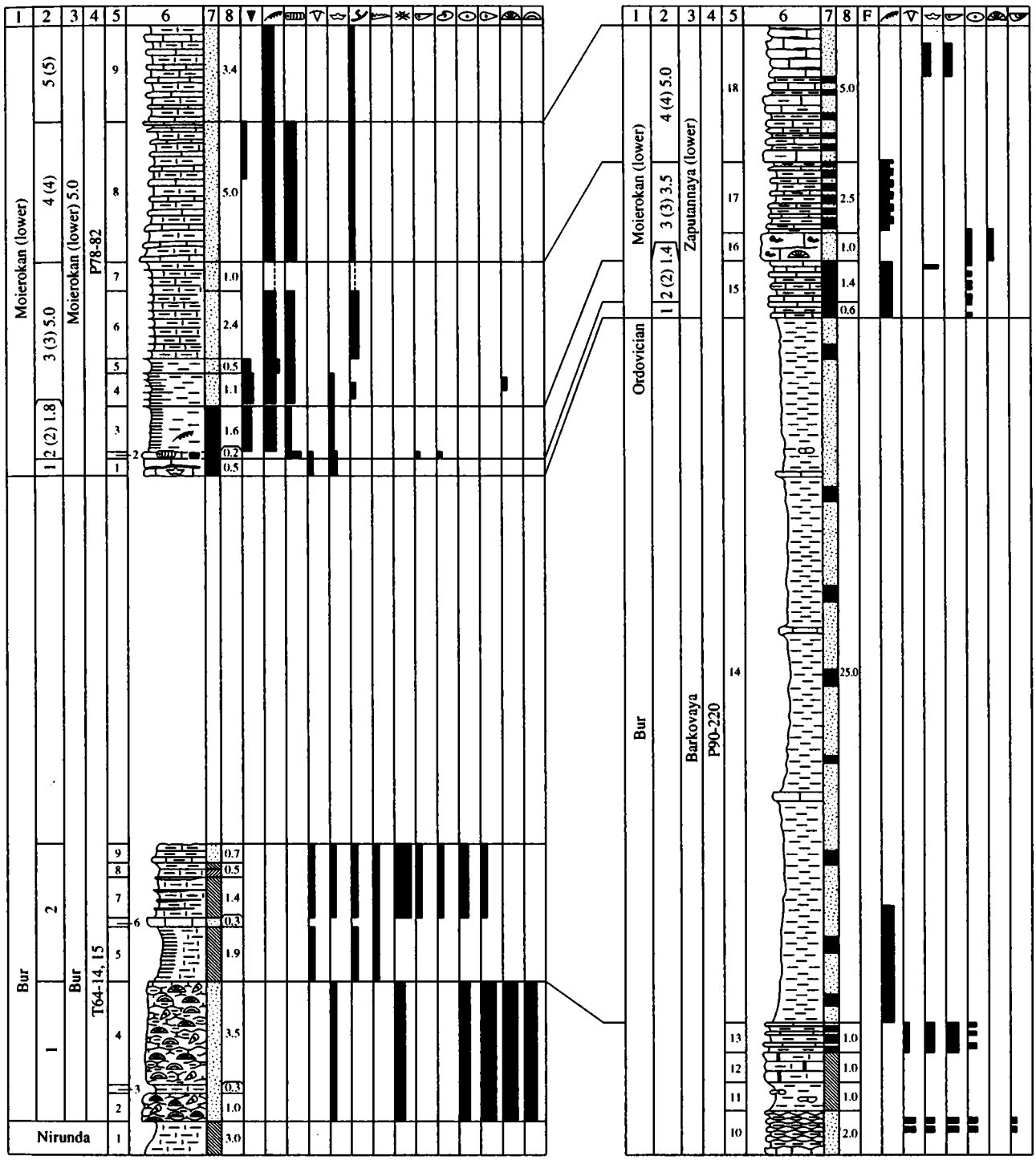


**Fig. 3.** (Contd.)

Beds 83/27 and 83/28: gray dolomitic grumous limestones with the tubercular lamination, abundant detritus, and frequent marl laminae. In the middle of Bed 83/27 and in boundary portions of Bed 83/28, nodular segregations are larger in size. Both beds enclose thin (1–2 cm) but frequent interlayers of nodular-detrital limestones. Small and abundant organic remains are buried in a dispersed or clustered state. Dominant forms are *Isorthis neocrassa* in the Bed 83/27 and *Zygospiraella duboisi* in the Bed 83/28 (13.0 m)..

*Khaastyr Horison*, hs, (129.5 m).

**Lower Khaastyr Subhorizon** (lower middle Aeronian), $hs_1$, (79.5 m).

(14) *Chronozone 1*, $hs_{1(1)}$: Beds 78/18–78/24 (20.3 m).

Beds 78/18–78/24: gray to brown-gray nodular limestones composed of tubercular microbeds (2–4 cm) separated by thin (3 mm and less) laminae of marls. These rocks incorporate abundant lens-shaped bodies of gastropod–brachiopod–bryozoan limestones. In the middle part of the unit (Bed 78/21), there are interlayers of detrital-nodular limestones displaying either normal, or cross lamination. Isolated fossils, as well as their clusters and accumulations are abundant. *Zigospiraella duboisi* (Bed 78/18), *Chasmotopora moyerensis* (Bed 78/19), *Loxoplocus vostokovae* (Bed 78/20), *Calamopora alveolaris* (Bed 78/21), and *Quadralites quadratus* (Beds 78/22–78/24) are dominant forms of fossil assemblages (20.3 m).

(15) *Chronozone 2*, $hs_{1(2)}$: Beds 78/25–78/31 and 90/18–90/21 (20.3 m).

Beds 78/25–78/31: gray grumous polydetrital limestones with tubercular lamination and interlayers (3–6 cm thick) of marls and bryozoan-crinoidal coquina. In the middle part (Bed 78/28), greenish gray clayey marls enclose frequent and thin (3 cm and less) interlayers of organogenic-detrital limestones with abundant mudeater tracks at the bedding planes. These planes also were the accumulation sites of abundant fossils. *Favosites gothlandicus gothlandicus* (Beds 78/25 and 78/26), *Stegerhynchus decemplicatus duplex* (Beds 78/27 and 78/28), *Eocoelia hemisphaerica* (Bed 78/29), and *Calamopora alveolaris* (Beds 78/30 and 78/31) are dominant forms of fossil assemblages (12.5 m).

Beds 90/18–90/21: greenish gray nodular limestones incorporating some detritus, small limy nodules, and frequent lenticular interlayers of bluish gray detrital limestones with flat autoclasts of the dark-gray clayey limestone. Bedding planes are tubercular. Small- and medium-sized organic remains are characteristic of limestones, where they occur in the dispersed and clustered state. The dominant form is *Eocoelia hemisphaerica* (7.9 m).

(16) *Chronozone 3*, $hs_{1(3)}$: Beds 90/22, 90/23, and 77/1–77/10 (18.0 m).

Bed 90/22: greenish gray grumous to nodular marls with rare thin lenses of gray detrital limestones. Abundant organic remains variable in size are scattered in the rocks. *Eocoelia hemisphaerica* is dominant (5.5 m).

Bed 90/23: gray grumous clayey limestones. The rocks enclose clay interlayers and dispersed small- and medium-sized fossils, among which *Favosites gothlandicus gothlandicus* is dominant (1.4 m).

Beds 77/1–77/10: gray grumous limestones with thin lenticular interlayers of organogenic limestones. The rocks are rich in variable detritus (up to 50%) and show bioturbation of the zoophycos type. Abundant organic remains occur in the isolated state, or as small accumulations. *Mesofavosites dualis* (Bed 77/1), *Zygospiraella duboisi* (Beds 77/2, 77/3, 77/5, and 77/6), *Eocoelia hemisphaerica* (Beds 77/4, 77/7, and 77/10) and *Favosites gothlandicus gothlandicus* (Bed 77/9) are dominant form of fossil assemblages (10.4 m).

(17) *Chronozone 4*, $hs_{1(4)}$: Beds 77/11–77/19 (21.0 m).

Beds 77/11–77/14: green friable calcareous shales with thin lenticular interlayers of gray organogenic polydetrital limestones. Small organic remains occur only in the latter. The dominant *Eocoelia hemisphaerica* occurs in coquina lenses and is widely scattered at the bedding planes (10.6 m).

Beds 77/15 and 77/16: gray grumous detrital limestones with bioturbation of the zoophycos type. Small- to medium-sized frequent organic remains are dispersed in the rocks. The dominant form is *Favosites gothlandicus gothlandicus* (2.7 m).

Beds 77/17–77/19: green friable shales with thin interlayers of detrital limestones. Small and rare organic remains occur mainly in limestones. *Eocoelia hemisphaerica* is dominant (7.5 m).

**Upper Khaastyr Subhorizon** (upper middle–upper Aeronian), $hs_2$, (50.0 m).

(18) *Chronozone 1*, $hs_{2(1)}$: Beds 77/20–77/26 (16.5 m).

Beds 77/20–77/25: gray clastic-polydetrital limestones bearing autoclasts, greenish gray dolomitic marls, and green friable shales are intercalated with lenticular interlayers (0.1 m) of grumous detrital limestones. Bioturbation of the zoophycos type and mudeater tracks are characteristic of some interlayers. Small- to medium-sized abundant fossils are buried on bedding planes or concentrated in lenses between layers. *Favosites gothlandicus gothlandicus* (Beds 77/20, 77/21, and 77/25), *Pentamerus oblongus* along with *Eocoelia hemisphaerica* (Beds 77/22 and 77/24), and *Mendacella tungussensis* (Bed 77/23) are dominant in fossil assemblages (9.9 m).

Bed 77/26: rhythmic alternation of layers composed of gray grumous polydetrital clayey limestone with rock packets consisting of interlaminated detrital limestones, green shales barren of fossils, and greenish gray marls. Small- to medium-sized fossils are concentrated on bedding planes of grumous limestones. *Eocoelia hemisphaerica* is dominant (6.6 m).

(19) *Chronozone 2*, $hs_{2(2)}$: Beds 77/27–77/35 (13.6 m).

Beds 77/27–77/30: gray clayey-dolomitic detrital limestones with pyrite. Thin lenticular interlayers of nodular limestones enclose laminae of yellow-green marls. Small- to medium-sized fossils are dispersed. *Favosites gothlandicus gothlandicus* is dominant. The rocks show bioturbation of the zoophycos type and large mudeater tracks. The upper Bed 77/30 is represented by the brachiopod coquina, where matrix consists of unsorted detritus mixed with fine-grained cal-

cite and dolomite. *Borealis borealis* is dominant at this level (5.4 m).

Beds 77/31–77/35: greenish gray marls and shales (interlayers up to 0.3 m thick) with flat nodules of gray limestone and thin laminae of detrital limestones bearing autoclasts. In the middle part (Bed 77/32 0.6 m thick), the rocks are gray, grumous, and bearing crinoids. Other fossils are buried in the dispersed or clustered state. *Eocoelia hemisphaerica* is dominant in marls and along bedding planes. Some limestone interlayers are bioturbated (7.3 m).

(20) *Chronozone 3*, $hs_{2(3)}$: Beds 77/36–77/43 (20.0 m).

Beds 77/36–77/38: gray clayey-silty-dolomitic detrital limestones with the grumous structure. Clay laminae separate individual limestone layers. *Favosites gothlandicus gothlandicus* is the dominant taxon of the unit (7.1 m).

Bed 77/39: gray to black platy grumous limestones enclosing corals and black laminae of shales. The upper part is foliated and dolomitic. The dominant taxon is represented by *Favosites gothlandicus gothlandicus* (4.7 m).

Beds 77/40–77/43: light-gray clayey-silty-dolomitic limestones with detrital-nodular phytolitic interlayers. Small- and medium-sized fossils are rare and dispersed. *Favosites gothlandicus gothlandicus* is dominant among them (8.3 m).

*Agidy Horizon*, ag, (44.5 m).

Lower Agidy Subhorizon (lower-lowermost middle Telychian), $ag_1$, 42.5 m.

(21) *Chronozone 1*, $ag_{1(1)}$: Beds 84, 85/8–85/22 (21.6).

Beds 84/8–84/12: gray, platy to foliated limestones horizontally laminated and alternating with dark-gray dolomitic detrital limestones enclosing laminae (0.2–0.4 cm) of greenish gray marls. Small and usually rare organic remains locally represent the rock-forming components, but are absent in other levels. *Herrmannina moierensis* (Beds 84/8, 84/9, and 84/11), *Lenatoechia multicostata* (Bed 84/10), and *Favosites gothlandicus gothlandicus* (Bed 84/12) are dominant taxa (8.2 m).

Beds 84, 85/13–85/18: alternating shales and marls green, brown, and variegated in color. The unit also incorporates lenticular interlayers of gray to lilac-red clayey-dolomitic stromatolite-detrital limestones. Small organic remains are irregularly dispersed and locally form small coquina lenses. *Cytherellina oviformis* (Bed 84/13), *Alispira rotundata* (Bed 84/14), *Herrmannina moierensis* (Beds 85/16 and 85/17), and *Anabaria rara* (Bed 85/18) represent the dominant taxa (6.7 m).

Beds 85/19 and 85/20: alternating platy and grumous limestones with pyrite, ripple marks, and occasional interlayers of shales. Small and abundant organic remains are irregularly dispersed. *Anabaria rara* (Bed 85/19) and *Herrmannina moierensis* (Bed 85/20) are dominant (4.0 m).

Beds 85/21 and 85/22: gray grumous clayey-detrital limestones displaying mudeater tracks and bioturbation of the zoophycos type. There are rare and thin marl interlayers in the beds. *Favosites gothlandicus gothlandicus* (Bed 85/21) and *Multisolenia tortuosa* (Bed 85/22) are dominant among abundant and scattered organic remains (2.6 m).

(22) *Chronozone 2*, $ag_{1(2)}$: Beds 86/23–86/34 (20.9 m).

Beds 86/23–86/34: green to lilac-brown (Bed 86/28b) and spotted (Bed 86/26) grumous shales with clots of calcareous limestones. Interlayers between their beds consist of dolomitic-calcareous marl, clayey-detrital limestone, and gray microdetrital limestone with autoclasts of green clayey marl and gray nodular limestone. Small organic remains are either dispersed, or occur in accumulations. *Cytherellina oviformis* (Beds 86/23–86/25), *Norilskinia cuspidata* (Beds 86/26–86/28), *Favosites gothlandicus gothlandicus* (Bed 86/29), *Cryptothyrella norilica* (Beds 86/30 and 86/31), *Mendacella tungussensis* (Bed 86/32), *Subalveolites subulosus* (Bed 86/33), and *Anabaria rara* (Bed 86/34) represent dominant taxa of the unit (20.9 m).

**Upper Agidy Subhorizon** (terminal middle–upper Telychian), $ag_2$, (42 m).

(23) *Chronozone 2*, $ag_{2(1)}$: Beds 87/35–87/40 (17.6 m).

Beds 87/35 and 87/36: gray detrital grumous limestones. Small- to medium-sized organic remains are usually dispersed and sometimes occur in clusters. *Multisolenia tortuosa* is dominant. Coral-stromatoporoid bioherms (1.2 × 1.5 m) are present in the unit (2.4 m).

Bed 87/37: greenish gray marls with clots of clayey limestone. Frequent small- to medium-sized fossils are dispersed. *Favosites gothlandicus moyeroensis* is the dominant taxon (2.2 m).

Beds 87/38–87/40: gray grumous clayey-detrital limestones. *Favosites gothlandicus moyeroensis* is dominant among fossils dispersed in the unit. Bioherms are present in the basal part of Bed 87/38, and Bed 87/40 incorporates platy limestones (6.7 m).

(23) *Chronozone 2*, $ag_{2(2)}$: Beds 87/41–87/52 (14.8 m).

Bed 87/41: brown-gray thin-platy limestone barren of organic remains and containing interlayers of green calcareous shale with flat authigenic pebbles of clayey limestone. Small organic remains occur in a dispersed state and in lentil accumulations. *Cytherellina oviformis* represents the dominant taxon (1.4 m).

Bed 87/42: brown-gray limestone with clayey micronodular laminae. Small dispersed organic remains include the dominant *Cytherellina oviformis* (2.0 m).

Beds 87/43–87/47: intercalation of brown-gray platy limestones and green friable shales with nodules and lenticular laminae of greenish gray clayey limestone. Small organic remains are dispersed and locally occur in small accumulations. Their dominant form is *Cytherellina oviformis* (4.4 m).

Bed 87/48: gray dolomitic limestone with wavy lamination, hieroglyphs, and sporadic laminae of detrital limestone. Ostracodes are dominant fossils of the bed (2.5 m).

Beds 87/49–87/52: intercalation of gray limestone layers, bearing crinoidal detritus and autoclasts, and greenish gray clayey marls enclosing nodules and lenses of gray fine-grained limestone. Hieroglyphs and mudeater tracks are abundant. *Alispira rotundata* is dominant among rare and small organic remains (4.5 m).

(25) *Chronozone 3*, $ag_{2(3)}$: Beds 87/53–87/56 (9.4 m).

Bed 87/53: gray platy dolomitic limestone ocherous at the surface and displaying ripple marks and mudeater tracks (4.1 m).

Beds 87/55 and 87/56a: gray to dark-gray grumous-platy microphytolitic limestones with rare ostracodes, crinoidal fragments, pelecypods, and hieroglyphs (3.4 m).

Bed 87/56b: greenish gray shales with rare interlayers (up to 1.5 cm thick) and lenses of limestone. The upper portion of the bed (0.4 m) is composed of platy dolomite. The bed yielded rare valves of ostracodes (1.9 m).

*Khakoma Horizon*, hk, (82 m).

**Lower Khakoma Subhorizon** (Sheinwoodian), $hk_1$, (42 m).

(26) *Chronozone 1*, $hk_{1(1)}$: Beds 60/3,60/4, 58A/2, and 58A/3 (9.2 m).

Beds 60/3 and 60/4: gray organogenic-detrital limestones brown at the surface. Small flat limestone pebbles and rounded stromatoporoids occur at the base of the unit, thus suggesting an erosion at the lower boundary of Bed 60/3. Abundant organic remains (biostrome) imply the auto- and allochthonous origin of burials. *Yavorskiina rzonsnickii* (Bed 60/3) and *Labechia condensa* (Bed 60/4) are dominant among fossils (2.6 m).

Bed 58A/2: gray grumous and thin-platy limestones. *Beyrichia mirabilis* is dominant in assemblage of rare small fossils (0.9 m).

Bed 58A/3: light-gray microdetrital sandy-silty limestone displaying platy to grumous structure and symmetrical ripple marks. *Herrmannina nana* is dominant among dispersed organic remains small or medium in size (2.0 m).

(27) *Chronozone 2*, $hk_{1(2)}$: Beds 58A/4–58A/8, 58/2–58/6 (8.3 m).

Bed 58A/4: stromatoporoid limestone grading upward into the stromatolitic variety. The rock matrix consists of gray detrital limestone with authigenic fragments. Interlayers in the bed are composed of gray thin-platy oncolitic limestone with sandy-silty admixture. *Ecclimadictyon fastigiatum* (level 58A/4b) and *Anabaria rara* (levels 58A/4e–58A/4f) are dominant taxa (1.6 m).

Bed 58A/5: yellow- to gray-green platy (1–3 cm) dolomitic marls with horizontal bedding (2.0 m).

Bed 58A/6: green friable shales frequently alternating with silty flat-laminated dolomites and limestones (1.4 m).

Bed 58A/7: intraformational conglomerate with large pebbles of silty, detrital, and oolitic limestones cemented by calcite. In the lower part of the bed, pebbles are oriented parallel to the bedding plane; at the top their orientation is tangential. *Beyrichia mirabilis* is dominant among rare fossils (0.5 m).

Bed 58A/8: greenish gray marls alternating with gray grumous cross-bedded limestones bearing an admixture of silty material. Abundant fossils are often concentrated in particular interlayers. *Beyrichia mirabilis* is dominant (1.7 m).

Beds 58/2–58/6: frequent intercalation of gray detrital grumous limestones and greenish gray clayey marls. Organic remains, rare or frequent, are concentrated in individual interlayers. The dominant form is *Beyrichia mirabilis* (1.1 m).

(28) *Chronozone 3*, $hk_{1(3)}$: Beds 58/3–58/15 (6.2 m).

Beds 58/7–58/14: gray, grumous to platy, detrital limestones with thin laminae of greenish gray marl. Stromatolite interlayers and Stromatoporata biostromes are present. Organic remains of small and medium size occur in the accumulated or dispersed state. *Beyrichia mirabilis* is dominant (4.8 m).

Bed 58/15: dark-gray grumous laminated limestone. Current marks are imprinted at the top surface (grooves are up to 5–8 cm deep). Stromatolites are present in the middle of the bed. Abundant fauna is regularly distributed throughout the bed, and only gastropods tend to be concentrated at the bedding planes. *Beyrichia mirabilis* and *Straparollus alacer* are dominant in the lower and upper layers, respectively (1.4 m).

(29) *Chronozone 4*, $hk_{1(4)}$: Beds 58/16–58/22 (8.5 m).

Beds 58/16–58/19: gray, irregularly bedded and grumous detrital limestones incorporating authigenic clasts and stylolites. Organic remains variable in size are buried at random or in accumulations. *Sapporipora favositoides* along with *Beyrichia mirabilis* (Bed 58/16), *Eohowellella yadrenkinae* (Bed 58/10), and *Prosolarium cirrhosa* (Bed 58/19) are dominant taxa (4.3 m).

Bed 58/20: banded alternation of greenish gray friable limestone and dolomite enclosing lenses of detrital limestones. The dominant taxon of the bed is *Bystrowicrinus bilobatus* (1.8 m).

Beds 58/21 and 58/22: thin-platy to foliated gray dolomites displaying exsiccation cracks and ripple marks at the bedding planes. Interlayers of gray clayey

marl are present. *Dalejina ribnayaensis* is dominant amid rare fossils (2.2 m).

(30) *Chronozone 5*, $hk_{1(5)}$: Beds 64/12–64/18 (9.8 m).

Beds 64/12–64/18: gray to dark-gray platy limestones with cherts, grumous limestone interlayers, and biostromes. Frequent organic remains are irregularly distributed. *Labechia condensa* is dominant; abundant *Neobeatricea nikiforovae*, *Ecclimadictyon fastigiatum*, and *Prosolarium cirrhosa* occur in some interlayers (9.8 m).

**Upper Khakoma Subhorizon** (Homerian), $hk_2$, (40.0 m).

(31) *Chronozone 1*, $hk_{2(1)}$: Beds 66/1b–66/6 (8.0 m).

Bed 66/1b: dark-gray thick-platy detrital limestone with stylolites; its upper part incorporates autobreccia. Organic remains are represented by crinoid fragments (0.7 m).

Bed 66/2: yellow-gray banded clayey limestone (0.1 m).

Beds 66/3–66/6: coral–stromatoporoid biostrome, where stromatoporoid laminae together with tabulate and rugose corals are cemented by dark-gray crinoidal limestone. Many colonies are overturned. *Megalomus* shells occur in the upper bed. The dominant taxon is *Yavorskiina membrosa* (7.2 m).

(32) *Chronozone 2*, $hk_{2(2)}$: Beds 66/7 and 66/8 (6.1 m).

Bed 66/7: yellow-gray cherty-clayey dolomite with rare clay interlayers. The rock is barren of fossils (0.75 m).

Bed 66/8: coral–stromatoporoid biostrome with laminated (at the base) and flat hemispherical (at the top) colonies of stromatoporoids, tabulates, and rugoses. Organic remains are cemented by yellow- to brown-gray clayey-detrital-dolomitic limestone. The dominant taxon is *Yavorskiina membrosa* (5.6 m).

(33) *Chronozone 3*, $hk_{2(3)}$: Beds 66/9–66/14 (5.1 m).

Beds 66/9–66/12: brown-gray, gray, and black platy clayey limestones with stromatolites and detrital material. The identified taxa are *Bystrowicrinus bilobatus* from Bed 66/10 and *Morinorhynchus proprius* from Bed 66/12 (2.3 m).

Bed 66/13: light- to brown-gray micronodular limestone with platy interlayers. The bed yielded only small shells of brachiopods (0.2 m).

Bed 66/14: brown-gray clayey-calcareous dolomite with vague platy structure. Fauna is absent (0.8 m).

(34) *Chronozone 4*, $hk_{2(4)}$: Beds 66/15–66/18 (4.8 m).

Beds 66/15–66/18: gray thin-platy biomorphic limestones. Organic remains medium in size are concentrated in particular layers (Beds 66/15, 66/17, and 66/18b), where *Clavidictyon cylindricum* is dominant (4.8 m).

(35) *Chronozone 5*, $hk_{2(5)}$: Beds 66/19–66/25 (8.2 m).

Beds 66/19 and 66/20: light-gray thin-platy clayey dolomite (Bed 66/19) grades upward into laminated limestone of the same color (Bed 66/20). The unit yielded only rare columnals of *Bystrowicrinus bilobatus* (1.2 m).

Beds 66/21–66/24: light- to brown-gray platy dolomites alternating with gray and dark-gray bituminous limy biostromes. Abundant organic remains medium in size are confined only to organogenic layers. Dominant forms are *Stelodictyon moierense* and *Clavidictyon cylindricum* (3.5 m).

Bed 66/25: light-gray thin-platy laminated limestone barren of fossils (3.0 m).

(36) *Chronozone 6*, $hk_{2(6)}$: Beds 66/26–66/31 (8.0 m).

Bed 66/26: black thick-platy limestone with abundant fossils dispersed or assembled in accumulations. *Stroparollus alacer* is dominant (1.2 m).

Bed 66/27: black micronodular limestones with small and rare organic remains The dominant taxon is *Beyrichia kureikiana* (0.6 m).

Beds 66/28–66/30: black to gray (Bed 66/30) platy limestones with dispersed small organic remains. *Beyrichia kureikiana* is dominant (3.2 m).

Bed 66/31: black interlayers of stromatolitic, coral-stromatoporoid, and detrital limestones. *Stelodictyon moierense* is dominant among organic remains medium to large in size, dispersed, or locally clustered (3.0 m).

*Tukal Horizon*, tk, (67.5 m).

**Lower Tukal Subhorizon** (basal lower Gorstian), $tk_1$, (27 m).

(37) *Chronozone 1*, $tk_{1(1)}$: Beds 11/3–11/13 (7.9 m).

Beds 11/3–11/10: alternation of gray and dark-gray medium- to thin-platy limestones represented by stromatolitic, detrital, brecciated, ostracod- or brachiopod-bearing microphytolitic, oolitic, micronodular, and stylolite-bearing varieties. Fossils are absent in some layers, but they are rock-forming components in others. *Murchisonia cingulata* (Bed 11/3), *Glebosites rihus* (Bed 11/7), *Eukloedenella kureikensis* (Bed 11/8), *Sibiricrinus helenae* (Bed 11/9), and *Favosites gothlandicus moyeroensis* (Bed 11/10) are dominant taxa of the unit (4.2 m).

Beds 11/11–11/13: dark-gray limy-dolomitic silty-detrital marls with thin-platy structure. The rocks are intercalated with gray lenses and thin interlayers of grumous, detrital, clayey, sandy-silty, and cross-laminated limestone varieties. Small and rare organic remains are concentrated at separate levels. The dominant taxon is *Eukloedenella kureikensis* (3.7 m).

(38) *Chronozone 2*, $tk_{1(2)}$: Beds 11/14–11/17 (6.4 m).

Beds 11/14–11/17: alternation of different rocks, such as hemispherical and columnar stromatolites, dark-gray and black oolitic limestones, dark-gray banded thin-platy shales, and thin-platy limestones.

*Bystrowicrinus bilobatus* (Bed 11/14) and *Hyattidina acutisummitatus* (Bed 11/17) are dominant among rare organic remains (6.4 m).

(39) *Chronozone 3*, $tk_{1(3)}$: Beds 3/2–3/13 (6.2 m).

Beds 3/2–3/12: gray marls with limy clots and gray to dark-gray limestones classed with thin-platy, detrital, clayey, oolitic, stromatolitic, and autoclast-bearing varieties. *Morinorhynchus proprius* (Bed 3/2), *Bazaricrinus parvulus* (Bed 3/3), *Hyattidina acutisummitatus* (Beds 3/4, 3/5, 3/11, and 3/12), *Eukloedenella kureikensis* (Beds 3/6, 3/10, and 3/11), *Monotrypa benjamini*, *Subalveolites subulosus* (Bed 3/7), and *Hyattidina parva* (Beds 3/8 and 3/9) are dominant taxa of diverse fossil assemblages (4.9 m).

Bed 3/13: dark-gray massive polydetrital-clayey limestone that yielded abundant fossils with dominant *Subalveolites subulosus* (1.1 m).

(40) *Chronozone 4*, $tk_{1(4)}$: Beds 3/14–3/28 (6.4 m).

Beds 3/14–3/23: thin-platy, dolomitic, clayey, and silty varieties of gray and dark-gray limestones, some of which enclose pyrite crystals, stylolites, silty lentils, laminae of calcareous clay, and interlayers of limy-dolomitic marls. Small organic remains are concentrated on bedding planes. *Straparollus alacer* (Beds 3/14–3/16) and *Hyattidina acutisummitatus* (Beds 3/17 and 3/18) are dominant taxa (3.4 m).

Bed 3/24: yellow-gray marl with moniliform inclusions and nodules of limestone. Rare lenticular interlayers (2–3 cm thick) of flat-laminated and brecciated limestones are present. *Hyattidina parva* is dominant among rare fossils (1.0 m).

Beds 3/25–3/27: dark-gray platy detrital-stromatolitic limestones. Small fossils are either rare (Beds 3/25 and 3/26) or abundant (Bed 3/27), and their dominant taxon is *Hyattidina acutisummitatus* (1.0 m).

Bed 3/28: stromatolites locally grading into detrital and detrital-oolitic limestones. Abundant shells of *Hyattidina acutisummitatus* are present in detrital limestones (1.0 m).

**Upper Tukal Subhorizon** (terminal lower–upper Gorstian), $tk_2$, (40.5 m).

(41) *Chronozone 1*, $tk_{2(1)}$: Beds 3/29, 3/30, and 2/4–2/15 (6.0 m).

Beds 3/29 and 3/30: greenish gray marl with abundant limy clots (Bed 3/29) and light-gray limestone (Bed 3/30) black in the upper part and bearing abundant specimens of *Eukloedenella kureikensis* (1.1 m).

Beds 2/4–2/15: greenish gray marls with clots and lenticles of detrital limestone and with interlayers (up to 0.5 m thick) of dark-gray massive polydetrital limestone, where small organic remains are dispersed and locally clustered. *Hyattidina acutisummitatus* (Beds 2/4–2/9, 2/14, and 2/15), *Beyrichia kureikiana* (Beds 2/10, 2/12, and 2/13), and *Bazaricrinus parvulus* (Bed 2/11) are dominant taxa (5.7 m).

(42) *Chronozone 2*, $tk_{2(2)}$: Beds 2/16–2/18 and 1/1–1/9 (6.0 m).

Beds 2/16–2/18: gray to dark-gray thin-platy detrital and micronodular limestones with a marl interlayer in the Bed 2/18. *Sibiricrinus helenae* (Beds 2/16 and 2/18) and *Bazaricrinus parvulus* (Bed 2/17) are dominant in their fossil assemblages (1.8 m).

Beds 1/1–1/9: platy, nodular, laminated, foliated, dolomitic, detrital, silty, and clayey varieties of limestones colored gray to dark-gray and intercalated with dark-gray marl interlayers. Biostromes with detrital limestone lenses are characteristic of Bed 1/8. *Bazaricrinus parvulus* (Bed 1/1), *Signetopsis cardinata* (Beds 1/2 and 1/3), *Sibiricrinus helenae* (Beds 1/4 and 1/7), *Beyrichia kureikiana* (Bed 1/5), *Amnestostroma fedorovi* (Bed 1/8), and *Leperditia lumaea* (Bed 1/9) represent dominant taxa (4.2 m).

(43) *Chronozone 3*, $tk_{2(3)}$: Beds 1/10–1/16, 4/2, and 4/3 (6.7 m).

Beds 1/10–1/14: gray micronodular limestone beds separated by limy-silty-dolomitic marl interlayers with lenses of gray detrital limestone and burials of ostracodes. *Sibiricrinus helenae* (Bed 1/10) and *Lioclema crustulum* (Beds 1/11–1/14) are dominant taxa of the unit (5.0 m).

Beds 1/15 and 1/16: clayey, detrital, and ostracode limestones colored gray, greenish gray, and black. Bedding planes are ornamented with exsiccation cracks. *Sibiricrinus helenae* is dominant among small organic remains (1.4 m).

Beds 4/2 and 4/3: gray to dark-gray platy and grumous clayey-detrital limestones with laminae (1–2 mm thick) of greenish gray marl. *Prosolarium cirrhosa* is dominant in assemblages of small and abundant fossils (1.3 m).

(44) *Chronozone 4*, $tk_{2(4)}$: Beds 4/4–4/12 (7.9 m).

Beds 4/4–4/7a: gray marls with abundant limy clots and interlayers of detrital-nodular limestone bearing autoclasts. *Prosolarium cirrhosa* (Bed 4/4), *Beyrichia kureikiana* (Bed 4/5), and *Healdianella inornata* (Bed 4/6) are dominant amid rare and dispersed fossils locally concentrated in lenticles (2.8).

Bed 4/7b: gray to greenish gray thin-platy dolomitic marl with detritus and fine banding (0.1–0.7 mm). Organic remains occur in the detrital fraction only (1.2 m).

Beds 4/8–4/12: greenish gray limy-dolomitic marls with conchoidal jointing and silty admixture; their beds alternate with green thin-platy clayey dolomites. *Bazaricrinus parvulus* (Bed 4/8), *Eridotrypa pervulipora* (Bed 4/9), *Sibiricrinus helenae* (4/10), *Bystrowicrinus quinquelobatus* (Bed 4/11), and *Densastroma astroites* (Bed 4/12) are dominant among organic remains either dispersed, or occurring in stratal accumulations (3.2 m).

(45) *Chronozone 5*, $tk_{2(5)}$: Beds 4/13 and 4/14 (5.1 m).

Bed 4/13: dark-gray massive grumous-polydetrital limestone; the massive limestone layers (0.5 m thick) are separated by thin (5 cm) interlayers of platy lime-

stone. Characteristic representatives of organic remains are large gastropods and fragmented Nautiloidea enveloped by mesophytolites. The dominant taxon is *Murchisonia cingulata* (2.5 m).

Bed 4/14: light-gray limy concretions (up to 2 × 4 cm) irregular in shape, composed of clayey-detrital limestone, and cemented by limy-dolomitic marl. *Eohowellella minimus* is dominant among fossils concentrated in the detrital fraction (2.6 m).

(46) *Chronozone 6*, $tk_{2(6)}$: Beds 10/28–10/40 (7.7 m).

Beds 10/28–10/32: detrital, massive, oolitic, bituminous, clayey, sandy-silty, and stromatolitic varieties of dark-gray to black limestones. *Leperditia lumaea* (Bed 10/28), *Murchisonia cingulata* (Bed 10/29), and *Parastriatopora kureikiana* (Beds 10/30–10/32) are dominant taxa of the unit (2.4 m).

Bed 10/33: dark-gray platy (up to 10 cm) to grumous limestones with abundant detritus. In the lower part, the bed encloses brown shale interlayers with abundant specimens of *Parastriatopora kureikiana* (0.9 m).

Beds 10/34–10/36: greenish gray fine-banded clayey limestones containing brecciated (Bed 10/34) and stromatolitic (Bed 10/36) interlayers. *Stigmatella indefinita* (Bed 10/34) and *Parastriatopora kureikiana* (Bed 10/36) represent species dominant in the rocks (2.3 m).

Bed 10/37: gray friable limy-dolomitic shales with rare fossils, among which *Herrmannina nana* is dominant (1.3 m).

Bed 10/38: dark-gray massive limestones with rare fossils, whose dominant representative is *Leperditia lumaea* (0.5 m).

Beds 10/39 and 10/40: greenish gray limy dolomites with horizontal bedding and isolated lenticles (1 × 5 cm) of gray polydetrital limestone. *Leperditia lumaea* is dominant amid rare organic remains (1.3 m).

*Postnichnyi Horizon*, ps, (81 m).

**Lower Postnichnyi Subhorizon** (Ludfordian), $ps_1$, (29.0 m).

(47) *Chronozone 1*, $ps_{1(1)}$: Beds TT-1/106–TT-1/111 (9.8 m).

Bed TT-1/106: white brecciated gypsum with domerite interlayers (2.5 m).

Bed TT-1/107: green fine-banded to nodular clayey domerite (0.9 m).

Bed TT-1/108: alternation of anhydrite–dolomite breccias and anhydrite layers with dark-gray banded domerites displaying hieroglyphs and small mudeater tracks (4.5 m).

Bed TT-1/110: gray massive or brecciated anhydrite layers with interlayers of gray clayey dolomites (0.8 m).

Bed TT-1/111: greenish gray massive to laminated clayey dolomites (1.1 m).

(48) *Chronozone 2*, $ps_{1(2)}$: Beds TT-1/112–TT-1/114a (11.2 m).

Bed TT-1/112–TT-1/114a: gray massive and locally brecciated anhydrite layers (10–20 cm thick) alternating with layers (15–20 cm thick) of gray banded domerite bearing sulfates (11.2 m).

(49) *Chronozone 3*, $ps_{1(3)}$: Beds TT-1/114b and TT-1/115 (7.9 m).

Bed TT-1/114b: gray to greenish gray massive dolomite with laminae of green domerite (2.5 m).

Bed TT-1/115: gray banded massive and brecciated anhydrite layers with laminae of greenish gray domerite (5.4 m).

**Middle Postnichnyi Subhorizon** (lower Pridoli), $ps_2$, (20.3 m).

(50) *Chronozone 1*, $ps_{2(1)}$: Beds TT-1/116–TT-1/119 (8.5 m).

Bed TT-1/116: gray banded gypsum-bearing domerite enclosing an interlayer (1 m thick) of vaguely laminated dolomite in the middle of the bed (2.8 m).

Bed TT-1/117–TT-1/119: gray dolomites with rugged laminae of black shales. In the bed TT-1/228, there is an anhydrite interlayer. Smooth ostracode valves are widespread in the unit (5.7 m).

(51) *Chronozone 2*, $ps_{2(2)}$: Beds TT-1/120–TT-1/124 (11.8 m).

Bed TT-1/120: gray banded dolomite with interlayers (10–30 cm thick) of anhydrite–dolomite breccia (3.0 m).

Bed TT-1/121 and TT-1/122: gray dolomite with two interlayers of gypsum–dolomite breccia and one interlayer of black banded shale (3.5 m).

Bed TT-1/123: gray brecciated gypsum with domerite interlayers gypsum-bearing and displaying the horizontal lamination (0.8 m).

Bed TT-1/124: gray banded clayey domerite with horizontal bedding (2.2 m).

(52) *Chronozone 3*, $ps_{2(3)}$: Beds TT-1/125–TT-1/128 (11.2 m).

Bed TT-1/125: gray massive and fine-banded dolomite. The bed yielded smooth ostracode valves and eurypterid fragments (5.4 m).

Bed TT-1/126–TT-1/128: gray to dark-gray banded domerite layers bearing anhydrite and displaying hieroglyphs. In the middle Bed TT-1/127, there is an interlayer of anhydrite–dolomite breccia (5.8 m).

**Upper Postnichnyi Subhorizon** (upper Pridoli), $ps_3$, (20.3 m).

(53) *Chronozone 1*, $ps_{3(1)}$: Beds TT-1/129–TT-1/135 (9.1 m).

Bed TT-1/129 and TT-1/130: gray brecciated anhydrite with interlayers of gray clayey domerite (2.5 m).

Bed TT-1/131: anhydrite–dolomite breccia and domerite interlayers with horizontal lamination and clots of anhydrite (1.4 m).

Bed TT-1/132–TT-1/135: gray domerites; in the middle (Bed TT-1/133), the rocks are sulfate-bearing and banded due to presence of anhydrite laminae (5.2 m).

(54) *Chronozone 2*, $ps_{3(2)}$: Beds TT-1/136–TT-1/140 (11.2 m).

Bed TT-1/136: gray wavy-laminated dolomite with gypsum inclusions and thin interlayers of authigenic breccia. The bed yielded rare bivalve remains (3.5 m).

Bed TT-1/137 and TT-1/138: gray banded domerites with five interlayers of gypsum–dolomite breccia and quartzite grains (4.5 m).

Bed TT-1/139: dark-gray laminated clayey dolomite with inclusions of anhydrite crystals and silty quartz grains. Fragmented shells of eurypterids are encountered (2.0 m).

Bed TT-1/140: gray wavy-laminated and sparsely-banded domerite (1.2 m).

The bed sequence is overlain by the Lower Devonian Yampakhta Formation composed of gray massive dolomites.

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# Evolution of Conodonts during the Late Cambrian and Early Ordovician

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**Abstract**—Proto-, para-, and euconodonts are characterized, and their stratigraphic significance is evaluated. General evolutionary trends and stages of development are depicted for the *Proconodontus*, *Teridontus*, and *Hirsutodontus* lineages of euconodonts. The potential of using each of the lineages for the terminal Upper Cambrian–Lower Ordovician conodont zonation is outlined for the cold-water, transitional, and warm-water paleobiogeographic provinces. The *Hirsutodontus* lineage, whose development commenced since the Late Cambrian (the Batyrbai Age commencement), is established to be independent of the *Teridontus* lineage. The pre-Trempealeauan interval of the *Teridontus* lineage development is proved to be characterized by a high evolution rate, and feasibility of using it for elaboration of the lower Upper Cambrian conodont zonation is suggested. It is also demonstrated that evolutionary transitions between euconodont taxa characteristic of the Batyrbai reference section (southern Kazakhstan) delimit the conodont zones of the section and are applicable for detailing the Upper Cambrian–Lower Ordovician conodont zonation in the transitional paleobiogeoraphic province as a whole. The conodont evolution is divided into five stages separated by distinct boundaries. The Cambrian–Ordovician boundary is shown to be marked by principal changes in the conodont evolution.

**Key words:** *conodonts, lineage, evolutionary stages, Late Cambrian, Early Ordovician*

## CURRENT UNDERSTANDING OF PROTO-, PARA-, AND EUCONODONT EVOLUTION, AND OBJECTIVES OF THE STUDY

The second half of the Late Cambrian, together with the initial Early Ordovician, represents the only period when proto-, para-, and euconodonts coexisted. The respective groups of fauna were different in the structure of elements composing the conodont apparatus, i.e., in the mode of lamellae growth (Bengtson, 1976, Fig. 11). Protoconodonts were the products of the outer epithelium. The growth of paraconodont lamellae occurred in the state when the lower parts of elements were immersed into the epithelial plica. The euconodont lamellae grew in the centrifugal manner, when the elements were entirely immersed into epithelium. The difference between these conodont groups seems to be sufficient to consider them as representing different orders. However, Clark *et al.* (1981) classed the proto- and paraconodonts with superfamilies Amphigeisinacea Miller and Furnishinacea Müller et Nogami, respectively, both belonging to the order Paraconodontida Müller, whereas the widespread and more diverse euconodonts represent, in their opinion, the order Conodontophorida Eichenberg. Müller and Hinz (1991) are of the same opinion, but they excluded the rank of superfamilies from consideration. Another classification was suggested by I.S. Barskov (see Maslakova *et al.*, 1995), but both the proto- and paraconodonts are included into the order Paraconodontida Müller even in this case.

Protoconodonts are known to occur from the basal Cambrian (Manykai Stage) to the Lower Ordovician and represent fossils appropriate for the zonal subdivision of the Lower Cambrian. Amid six protoconodont genera, such as *Protohertzina* Missarzhevsky, *Amphigeisina* Bengtson, *Hertzina* Müller, *Gapparodus* Abaimova, *Gumella* Müller et Hinz, and *Phakelodus* Miller, only the latter survived until the Early Ordovician, the early Arenigian Age included. So far, a possible evolutionary transition from proto- to paraconodonts is only suggested in theory, because concrete fossils indicative of such an event are still unknown (Müller and Hinz-Schallreuter, 1996). The event presumably happened close to the Lower–Middle Cambrian boundary (Bengtson, 1983, Fig. 7; Miller, 1984, Fig. 2), as the paraconodonts first appeared in the Middle Cambrian. They developed until the end of the Tremadocian and coexisted up to the middle Late Cambrian with the transient protoconodonts, thus being usable in the stratigraphic subdivision and correlation of the middle–lower Upper Cambrian. This is evident, for instance, from the sole paraconodont zonation elaborated in northern China for the Middle–Upper Cambrian (An, 1982).

Two lineages are distinguished in the paraconodont evolution (Miller, 1984). The first one includes the V- and W-shaped elements attributed to the genus *Wester-*

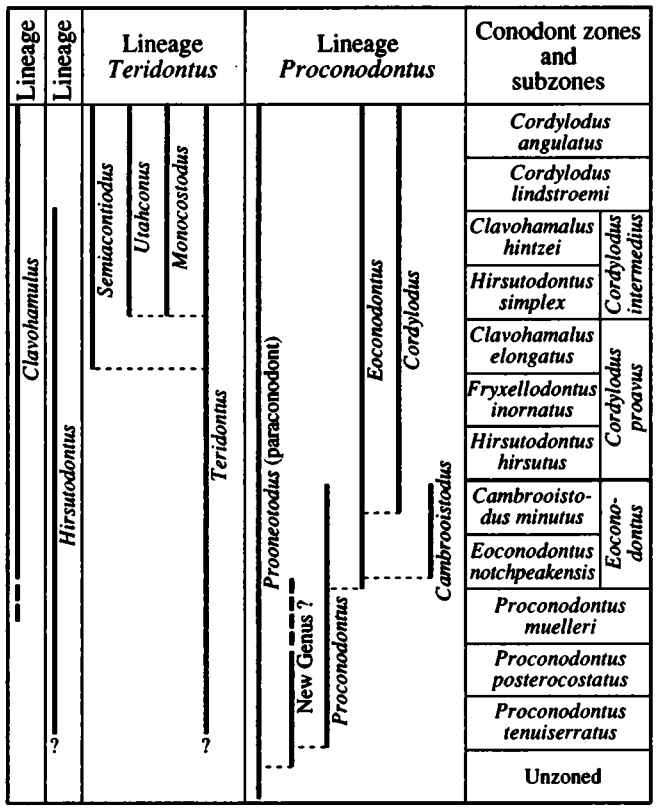
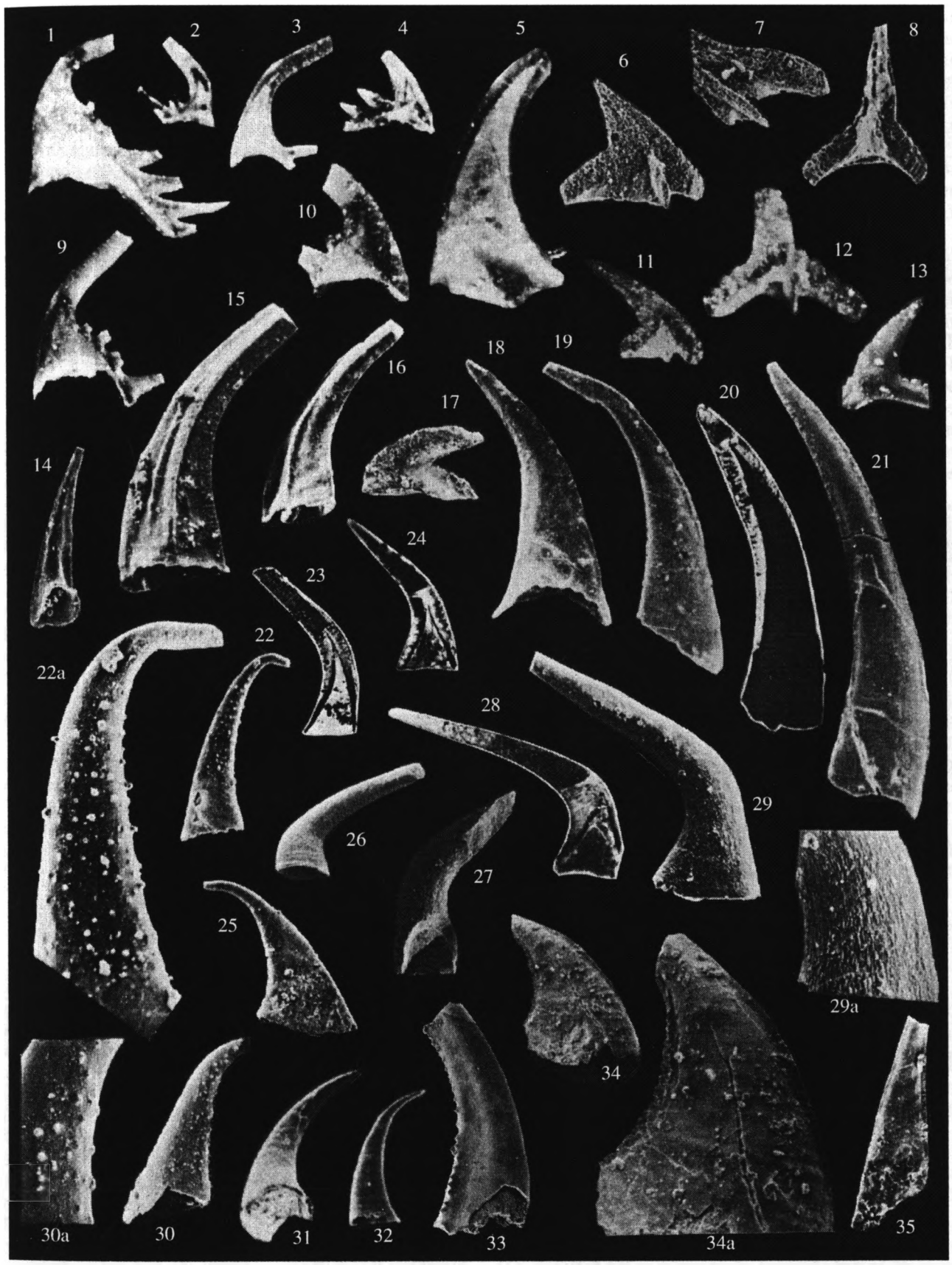



**Fig. 1.** Evolution of Late Cambrian and Early Ordovician conodonts according to Miller (1984) with modifications in the *Teridontus* and *Hirsutodontus* lineages.

*gaardodina* Müller. The second lineage is represented by simple cone conodonts of genera *Furnishina* Müller, *Muellerodus* Miller, *Prooneotodus* Müller et Nogami, *Prosagittodontus* Müller et Nogami, *Proacodus* Müller, *Bengtsonella* Müller et Hinz, *Trolmenia* Müller et Hinz, and others. Relations between taxa of both lineages are unclear, although evolutionary transitions from both to euconodont lineages appear to be possible (Miller, 1980, 1984). Miller (1984) substantiated the general conclusion of Bengtson (1976) that paraconodonts were ancestral with respect to the euconodont elements. He distinguished the *Chosonodina*, *Fryxellodontus*, *Teridontus*, and *Proconodontus* lineages in the euconodont evolution. It was suggested that various species of two paraconodont lineages might be ancestral for one or two of euconodont lineages. For instance, it is probable that euconodonts of the *Chosonodina* lineage originated from *Westergaardodina* paraconodonts at the beginning of the late Tremadocian (the appearance time of first *Chosonodina* Müller). The origin period of *Fryxellodontus* Miller and affinity between this genus and paraconodonts are not very well known, but the first primitive euconodonts of *Teridontus* and *Proconodontus* genera may be quite definitely affiliated with the evolution of paraconodonts. Both of them appeared in the middle Late Cambrian. The exact ancestor of primitive *Teridontus* forms is unknown, though they seem to be descendants of a conical taxon from the second paraconodont lineage of Miller. *Proconodontus tenuiserratus* Miller, the most primitive euconodont of the genus, is shown to be an offspring of paraconodont *Prooneotodus rotundatus* (Druce et Jones). It was also suggested that the basal cone of the euconodont element is homologic to the whole paraconodont element (Miller, 1980). Actually, the basal cone of *P. tenuiserratus* Miller is almost identical to the whole element of *P. rotundatus* (Druce et Jones), although the former has a newly formed thin layer of

**Plate.** Some conodont specimens from the Batyrbai reference section of southern Kazakhstan:

(1–3) *Cordylodus proavus* Müller, rounded (1 and 3) and compressed (2) elements, ×70, specimen nos. 4725/147 from level of 152 m(1), 4725/157 from level 152 m (2), 4725/148 from level 129.5 m (3). (4, 10) *Cordylodus intermedius* Furnish, lateral view, ×70: (4) compressed element, specimen no. 4725/160 from level of 205 m; (10) rounded element, specimen no. 4725/163 from level of 320 m. (5) *Cordylodus primitivus* Bagnoli, Barnes et Stevens, lateral view, ×70, specimen no. 4725/145 from level of 109 m. (9) *Cordylodus lindstromi* Druce et Jones, rounded element, lateral view, ×40, specimen no. 4725/165 from level of 221 m. (6–8, 11) *Prioniodus deltatus longibasis* (Mc Tavish) specimens from level of 625 m: (6, 11) prioniodiform elements, lateral views, specimen no. 4725/275, ×71 (6) and specimen no. 4725/279, ×40; (7) gothodiform element, lateral view, specimen no. 4725/290, ×74; (8) trichonodelliform element, posterior view, specimen no. 4725/293, ×68. (12, 13) *Prioniodus elegans* Pander, prioniodiform elements from level of 700 m, lateral views: (12) specimen no. 4725/348, ×70; (13) specimen no. 4725/347, ×40. (14–16) *Eoconodontus alisonae* Landing: (14) scandodiform element, posterior view, specimen no. 4725/129 from level of 93 m, ×120; (15, 16) drepanodiform elements, lateral views, ×200, specimen no. 4725/133 from level of 94.5 m (15) and specimen no. 4725/136 from level of 93 m. (17) *Cambrooistodus minutus* (Miller), oistodiform element, lateral view, ×60, specimen no. 4725/239 from level of 102 m. (18–21, 23, 24) *Eoconodontus notchpeakensis* (Miller), lateral views: (18, 24) scandodiform elements, specimen no. 4725/113 from level of 80 m, ×120, and specimen no. 4725/116 from level of 72.5 m displaying the white matter distribution, ×125; (19–21, 23) drepanodiform elements, specimen no. 4725/111 from level of 71 m, ×200 (19), specimen no. 4725/101 from level of 72.5 m displaying the white matter distribution, ×200 (20), specimen no. 4725/102 from level of 71 m, ×200 (21), and specimen no. 4725/123 from level of 72.5 m, ×125 (23). (22, 22a, 25) *Hirsutodontus transmutatus* (Xu et Xiang) from level of 24 m, lateral views: (22, 22a) specimen no. 4725/218, ×90 and ×240; (25) specimen no. 4725/219, ×80. (26–29a) *Teridontus nakamurai* (Nogami), lateral views: (26)specimen no. 4725/95 from level of 72.5 m, ×120; (27) specimen no. 4725/96 from level of 80 m, ×120; (28) specimen no. 4725/87 from level of 72.5 m displaying the white matter distribution, ×125; (29, 29a) specimen no. 4725/98 from level of 72.5 m, ×120 and ×300. (30, 30a) *Hirsutodontus ani* (Wang), lateral views, specimen no. 4725/93 from level of 42 m: (30) ×70; (30a) a part of the lateral surface, ×280. (31, 32) *Prooneotodus rotundatus* (Druce et Jones): (31) specimen no. 4725/85 from level of 2.5 m, inferior view, ×120; (32) specimen no. 4725/86 from level of 35 m, lateral view, ×70. (33) *Proconodontus tenuiserratus* Miller, lateral view, specimen no. 4725/398 from level of 8 m, ×61. (34, 34a) *Proconodontus serratus* Miller, lateral views, specimen no. 4725/99 from level of 64 m, ×70 and ×120. (35) *Proconodontus muelleri* Miller, lateral view, specimen no. 4725/232 from level of 71 m, ×76.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the euconodont histology enveloping the paraconodont basal cone. In this connection, Miller (1984) argued that the respective problem of phylogenetic relations should be solved after a detailed histologic investigation of elements identified as "gen. et sp. nov.?" and displaying morphology transitional between *Prooneotodus rotundatus* (Druce et Jones) and *Proconodontus tenuiserratus* Miller. The same problem can be investigated by using the most primitive representatives of the *Teridontus* lineage, which are less evolved than *T. nakamurai* (Nogami), and also the forms close to *Hirsutodontus ani* (Wang), which represent the oldest species of the genus. In addition, the primitive *H. ani* (Wang) and *H. transmutatus* (Xu et Xiang) from the middle Upper Cambrian deposits of Kazakhstan and China could be ancestors of more evolved *H. hirsutus* Miller, *H. simplex* (Druce et Jones), and some other species. In this case, the *Hirsutodontus* lineage could be an independent one in the euconodont evolution, thus excluding the phylogenetic affinity between genera *Hirsutodontus* Miller and *Teridontus* Miller suggested by their author (Fig. 1).

It is established at present that the oldest euconodonts *Proconodontus tenuiserratus* Miller (*Proconodontus* lineage), *Teridontus nakamurai* (Nogami), and *Hirsutodontus ani* (Wang) representing the proper lineages appeared almost concurrently in the middle Late Cambrian. They are known from North America (upper Franconian), Australia (pre-Peyntonian time), northern China (Chenshanian–Fengshanian boundary level), and Kazakhstan (the base of Batyrbai Stage or *Hirsutodontus ani* Zone). Accordingly, the middle Late Cambrian marks a very important event in the conodont evolution, i.e., the first occurrence level of representatives from the order Conodontophorida Eichenberg. Therefore, it is extremely important to elucidate general trends of three principal euconodont lineages and to evaluate their potential for elaboration of the conodont zonation. The problem is especially interesting in the aspect of paleobiogeoraphic factors.

In this work, I consider all data on conodonts from the reference Batyrbai section of southern Kazakhstan described in several publications (Chugaeva *et al.*, 1989; Dubinina, 1990; and others). Distribution patterns of conodonts are characterized in my dissertation (Dubinina, 1990). The fossils are pictured in the presented Plate and can be also found in the work by Apollonov *et al.* (1984).

## *PROCONODONTUS* LINEAGE EVOLUTION

The lineage includes genera *Proconodontus* Miller, *Eoconodontus* Miller, *Cambrooistodus* Miller, and *Cordylodus* Pander.

Morphogenesis of simple, smooth at the surface, conical elements of this lineage was effective, and a series of morphological changes was as follows: (1) thickening of the euconodont crown and parallel thinning of the basal cone having the paraconodont histology; (2) development of sculptural elements (keels, costae, serration of the posterior keel) characteristic of species from the genus *Proconodontus* Miller; (3) appearance of white matter in the euconodont crown and a simultaneous depth decrease in the basal cavity established in *Eoconodontus notchpeakensis* (Miller); (4) development of posterior process at the cone (species from the genus *Cambrooistodus* Miller); (5) emergence of denticles at the posterior edge of the conical element (*Cordylodus primitivus* Bagnoli, Barnes et Stevens); (6) formation of denticulate posterior process (*Cordylodus andresi* Viira et Sergeeva); (7) further evolution of cordylodiform elements, such as concentration and compaction of white matter, a depth decrease of basal cavities, morphological changes in basal cones, etc.

This trend in the euconodont evolution led to the apparatus complication. For instance, species from genus *Proconodontus* Miller (*P. tenuiserratus* Miller, *P. posterocostatus* Miller, *P. muelleri* Miller, *P. serratus* Miller) show thin euconodont crowns lacking white matter and distinct basal cones with clear transverse striation of paraconodont lamellae. These elements are grouped into a one-element apparatus. A little later (the middle of the Batyrbai Stage, or the middle Trempealeauan in North America), the origin of two-element apparatuses with rounded and flattened elements is recorded. Basal cavities of little depth and white matter in the cusp are characteristic of both elements. *Eoconodontus notchpeakensis* (Miller) is an example of this situation. Species of genus *Cambrooistodus* Miller, along with *Eoconodontus alisonae* Landing, appeared in the upper part of the Trempealeauan and Batyrbai stages. *E. notchpeakensis* (Miller) was the ancestral form for these taxa and also for species of genus *Cordylodus* Miller, whose most primitive representative is *C. primitivus* Bagnoli, Barnes et Stevens. In general, the Batyrbai locality in southern Kazakhstan (Apollonov *et al.*, 1984, 1985; Chugaeva *et al.*, 1989; Dubinina, 1991) and sections of western Newfoundland (Barnes, 1985, 1988; Bagnoli *et al.*, 1987) revealed two evolutionary lineages: *E. notchpeakensis–E. alisonae* and *E. notchpeakensis–Cordylodus primitivus* (Fig. 2).

*C. primitivus* Bagnoli, Barnes et Stevens (an equivalent of *Cordylodus* sp.) was first discovered in the Batyrbai section (Apollonov *et al.*, 1984, 1985) and then (an equivalent of *C. primitivus*) in western Newfoundland (Bagnoli *et al.*, 1987). It represents a form transitional between *E. notchpeakensis* (Miller) and real *Cordylodus* species; e.g., *Cordylodus proavus* Müller. *C. andresi* Viira et Sergeeva is another primitive form of the last genus that was presumably derived from *Proconodontus serratus* Miller (Szaniawski and Bengtson, 1993). This form is known from sections of Sweden, Estonia, the Leningrad district (Andres, 1981; Kaljo *et al.*, 1986; Popov *et al.*, 1989), and western Newfoundland (Barnes, 1988). *C. primitivus* Bagnoli, Barnes et Stevens and *C. andresi* Viira et Sergeeva are

still lacking the distinctive feature of the genus: the posterior denticulate process is underdeveloped in the former, and the density of the white matter is low in the latter (in its main cusp and in denticles of the posterior process).

A high rate of evolution that resulted in the appearance of four Late Cambrian genera and nine species, except for transitional forms, is a distinctive feature of the *Proconodontus* lineage, enabling the recognition of two stages in its development.

The first (Late Cambrian) stage resulted in the appearance of primitive apparatuses consisting of conical euconodonts with smooth walls and was characterized by a quick transition to the two-element apparatuses including the complex denticulate forms, which were lacking, however, the distinctive features of genus *Cordylodus* Pander.

During the second (Early Ordovician) stage, the evolution of *Proconodontus* lineage slowed down. As a result, a series of typical Tremadocian codylodids (*C. proavus* Müller–*C. intermedius* Furnish–*C. lindstromi* Druce et Jones–*C. angulatus* Pander–*C. rotundatus* Pander) was formed. Apparatuses of these species display a combination of rounded (cordylodiform) and compressed (cyrtoniodiform) elements.

Euconodonts of the *Proconodontus* lineage developed in three paleobiological provinces: cold-water, transitional, and warm-water (Dubinina, 1991). Unfortunately, only the most primitive species of genus *Proconodontus* Miller were inhibited in the transitional area and missing in the cold-water province. This situation slightly restricts the possibility of using the group in stratigraphic subdivision of the upper Upper Cambrian deposits.

## *TERIDONTUS* LINEAGE EVOLUTION

The lineage is represented by genera *Teridontus* Miller, *Semiacontiodus* Miller, *Monocostodus* Miller, *Utahconus* Miller, and *Rossodus* Repetski et Ethington.

In this work, I consider the *Teridontus* lineage as independent of the *Hirsutodontus* line (Fig. 1). All its representatives display microstriae on the surface of elements that is parallel to the axis of the main cusp; this makes them different from smooth elements characteristic of the *Proconodontus* lineage.

The *Teridontus* line evolution is again divisible into two stages. However, these stages, as compared with those of the previous lineage, lasted longer and proceeded in an opposite mode with respect to the rate of fauna evolution.

The first (Late Cambrian–earliest Ordovician) stage was characterized by a very slow development of the genus *Teridontus* Miller. The genus was represented solely by sole species *T. nakamurai* (Nogami) slightly variable in morphology and typical of Asian and Australian regions.

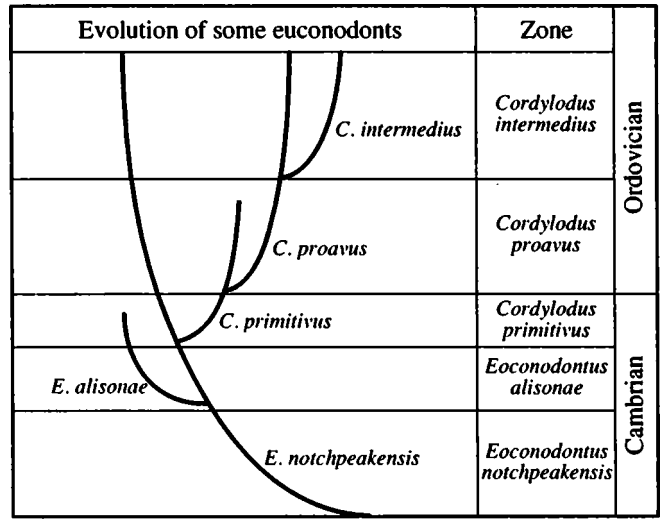


**Fig. 2.** Evolution lineages of some euconodonts during the Cambrian–Ordovician boundary interval (data from the Batyrbai reference section of southern Kazakhstan).

The second (Early Ordovician) stage (since the appearance time of *Cordylodus intermedius* until the Tremadocian, inclusive) marks the origin and swift evolution of species having apparatuses with costate conical elements and belonging to genera *Semiacontiodus* Miller, *Monocostodus* Miller, *Utahconus* Miller (early Tremadocian taxa), and *Rossodus* Repetski et Ethington (late Tremadocian taxon). All these forms inhabited the warm epicontinental seas of North America, northern China, and Australia. In other deep and cold basins, the *Teridontus* lineage was represented by the aforementioned species *T. nakamurai* (Nogami) that existed without noticeable morphological changes during the Late Cambrian and early Tremadocian.

Elements of the most ancient *T. nakamurai* (Nogami) already had the well-developed euconodontid crown and white matter in the cusp. These features and the low basal cavity combined with the relatively thin paraconodontid cone suggest that most primitive representatives of the lineage appeared in Asia and Australia during the Dresbachian or early Franconian time (Miller, 1980). Exactly this interval seems to be appropriate for seeking a transition from para- to euconodonts in the *Teridontus* lineage. In addition, the rate of the pre-Trempealeauan evolution of the lineage was likely rather high, and respective forms appear to be useful for zonal stratigraphy of the lower Upper Cambrian. Accordingly, we should attentively study the conodonts of this time interval.

## *HIRSUTODONTUS* LINEAGE EVOLUTION

In my opinion, this lineage evolved independent of the *Teridontus* line (Fig. 1). It includes one genus *Hirsutodontus* Miller, whose representatives had elements decorated with tubercles and processes at the surface.

We incompletely know the Late Cambrian interval of the lineage evolution. However, it is clear at present that its oldest forms, such as *H. ani* (Wang) and *H. transmutatus* (Xu et Xiang), were ancestors of the Early Ordovician species *H. hirsutus* Miller and *H. simplex* (Druce et Jones). An active speciation of genus *Hirsutodontus* took place during the Early Ordovician in the epicontinental seas of North America, northern China, and Australia, i.e., in the warm-water paleobiological province parental for *Semiacontiodus*, *Monocostodus*, and *Utahconus* forms of the *Teridontus* lineage (Dubinina, 1991). The indicated species were inhibited in the transitional zone and absent in the cold-water basins.

The oldest forms of the lineage in question, like *H. ani* (Wang), have primitive elements (a high basal cavity, the relatively thin euconodontid crown, and the massive paraconodontid basal cone), on the one hand, and the microcusp incorporating white matter, on the other (images 30 and 30a in the Plate). Taking into account these features, one may expect that there were more primitive forms absolutely lacking white matter. These forms, when found and investigated, may characterize a transition from para- to euconodonts in the *Hirsutodontus* lineage. In this case, the euconodont evolution would be shown corresponding to three or more independent lineages.

## DIFFERENCES IN EVOLUTION OF *PROCONODONTUS*, *TERIDONTUS*, AND *HIRSUTODONTUS* LINEAGES

Most primitive forms of the three euconodont lineages (*Proconodontus*, *Teridontus*, and *Hirsutodontus*) appeared almost concurrently in the middle Late Cambrian. They were, however, not identical with respect to the state of their evolution. *T. nakamurai*, the first representative of the *Teridontus* lineage, was most progressive among others, *H. ani* (Wang) of the *Hirsutodontus* lineage was at an intermediate state, and *P. tenuiserratus* Miller that represented the *Proconodontus* line was definitely most primitive.

The difference between the evolutionary rates of the three lineages is also obvious. In the Late Cambrian, the progressive species of the *Teridontus* lineage turned out to be absolutely "passive," the *Hirsutodontus* line was slightly "active," and the most primitive *Proconodontus* forms demonstrated a high rate of morphological changes. Apparatuses of the latter, originally consisting of simple conical elements, evolved up to the state of first complex elements of an imperfect codylodiform type. In the Early Ordovician, the lineages developed in a different manner. Evolution of *Proconodontus* species slowed down, and their morphological changes were related only to the further complication of cordylodiform elements. At the same time, representatives of *Hirsutodontus* and *Teridontus* lineages went through an active stage of evolution in warm epicontinental basins (see above).

Thus, the *Proconodontus* lineage appears to be the most important for subdivision and correlation of the Cambrian–Ordovician boundary deposits. Its species, except the most primitive *Proconodontus* forms missing in the cold basins, evolved inside three paleobiogeographic provinces. The *Teridontus* and *Hirsutodontus* lineages are useful in this aspect for the Lower Ordovician deposits of warm and transitional (to a lesser extent) paleobiogeographic zones. The last one can be also used for elaborating the conodont zonation in the uppermost Cambrian deposits of these zones.

## NEW DATA ON THE EARLY ARENIGIAN CONODONT EVOLUTION

It is unclear at present how the late Tremadocian and early Arenigian conodonts are interrelated with the described euconodont lineages. Considering the apparatus types, but excluding from consideration other parameters, e.g., the microsculpture of elements, Miller (1984) suggested that the late Tremadocian genera *Drepanoistodus* Lindstrom and *Drepanodus* Pander are descendants of *Cambrooistodus* Miller and *Eoconodontus* Miller of the *Proconodontus* lineage. Serious arguments in favor of this opinion are unknown, and my data on conodonts from the Batyrbai section of southern Kazakhstan do not help to solve the problem. However, this section demonstrated that there was the lineage *Prioniodus deltatus longibasis–Prioniodus elegans* of the early Arenigian time (Dubinina, 1991). Mc Tavish (1973), who described *Acodus deltatus longibasis* from the Emanuel Formation of West Australia, was first to assume the phyletic relations between the species of the lineage. Unfortunately, *Prioniodus deltatus longibasis* is missing in the Emanuel Formation, and relations between the ancestral form and descendants remained doubtful. In contrast, an interval with *Prioniodus deltatus longibasis* (Mc Tavish) was distinguished below the interval of *Prioniodus elegans* Pander, and the phyletic connection between both taxa are substantiated by the similarity of their apparatuses. It is important in this case that the apparatus of *Prioniodus deltatus longibasis* consists of more primitive prioniodiform and ramiform elements differentiating this species from *Prioniodus elegans* Pander, whose posterior (and some lateral) processes demonstrate an initial dentelation.

Pohler *et al.* (1987) reported the new form *Prioniodus* n. sp. discovered in the slope facies of western Newfoundland, which is similar in the apparatus structure to *Prioniodus elegans* Pander, although the dentelation of its elements is at a more primitive stage of development. They suggested that *Prioniodus* n. sp. was the ancestor of *Prioniodus elegans* Pander (Pohler *et al.*, 1987). I cannot exclude the possibility that *P.* n.

sp. from the western Newfoundland and *P. deltatus longibasis* (Mc Tavish) from the Batyrbai locality are closely related forms. If it is possible to prove the specific identity of both forms, then the lineage *Prioniodus deltatus longibasis–Prioniodus elegans* will be useful for detailing the conodont zonation of the lower Arenigian in the transitional paleobiogeographic province.

According to the latest data, *Prioniodus* n. sp. (Pohler *et al.*, 1987), now classed with *Prioniodus serratus* Pohler (Pohler, 1994), is evolved a little more than *Prioniodus deltatus longibasis* (Mc Tavish). At the same time, it impossible to exclude the chance that the former is the younger synonym of the latter, because differences between these species are insignificant.

## BATYRBAI SECTION AS AN EXAMPLE OF USING EUCONODONTS IN UPPER CAMBRIAN–LOWER ORDOVICIAN STRATIGRAPHY

The studied conodont assemblages from the Batyrbai section revealed a series of successive transitions between euconodonts. This situation was used to refine and complete the phylogenetic lineages of the latter characteristic of the Late Cambrian–Tremadocian, and to distinguish a new lineage that appeared in the early Arenigian.

(I) It is substantiated that the *Hirsutodontus* lineage of euconodonts developed independently since the Late Cambrian (initial Batyrbai Stage). The established morphological transition from *Hirsutodontus ani* to *Hirsutodontus transmutatus* proves this conclusion, although genera *Hirsutodontus* Miller and *Teridontus* Miller have been formerly considered as representing a single phylogenetic lineage (Miller, 1980, 1984).

(II) Phyletic relations in the Late Cambrian (terminal Batyrbai time) *Proconodontus* lineage are substantiated for *Eoconodontus notchpeakensis* (Miller) and *Eoconodontus alisonae* Landing, on the one hand, and for *E. notchpeakensis* (Miller) and *Cordylodus primitivus* Bagnoli, Barnes et Stevens, on the other. Both biotic events are recorded only in the transitional paleobiogeographic province, where they can be used to distinguish the *alisonae* and *primitivus* conodont zones and to refine the zonal scale of the Upper Cambrian.

(III) The early Arenigian lineage *Prioniodus deltatus longibasis–Prioniodus elegans* originally detected in the Batyrbai section can be important for the transitional province as a whole.

The detected changes in the euconodont phylogenetic lineages are used to delimit the *Hirsutodontus ani*, *Eoconodontus notchpeakensis*, *Eoconodontus alisonae*, *Cordylodus primitivus*, *Prioniodus deltatus longibasis*, and *Prioniodus elegans* zones in the Batyrbai section. This conodont zonation is one of few stratigraphic scales established in the transitional paleobiogeographic province.

## PRINCIPAL STAGES IN THE LATE CAMBRIAN–EARLY ORDOVICIAN CONODONT EVOLUTION

Analyzing changes in phyletic lines of euconodonts and the distribution patterns of proto- and paraconodonts, I have made an attempt to outline principal stages in their evolution during the Late Cambrian and Early Ordovician. Data available for different paleobiogeographic provinces impart a global character to these stages and boundaries separating them. Using this approach, I have recognized the late Aksai, Batyrbai, early–late Tremadocian, and early Arenigian stages of the conodont development.

(I) The late Aksai stage corresponded to the existence time of *Westergaardodina amplicava*. Fossil assemblages of the stage consist of proto- and paraconodonts exclusively.

(II) The Batyrbai stage characterizes the evolution time of *Hirsutodontus ani–Cordylodus primitivus* lineage. First euconodonts of *Proconodontus*, *Hirsutodontus*, and *Teridontus* genera appeared at this stage. In the *Proconodontus* lineage, forms with apparatuses consisting of simple elements were replaced by species possessing the first complex apparatuses. In the *Hirsutodontus* lineage, apparatuses changed insignificantly, and no change was recorded in the *Teridontus* lineage that included only one species at that time.

The stage is divisible into four substages.

(IIa) The *Hirsutodontus ani* time marks the first occurrence level of fossils representing the order Conodontophorida Eichenberg. These forms from genera *Hirsutodontus* Miller, *Teridontus* Miller, and *Proconodontus* Miller had one-element apparatuses of the euconodont type.

(IIb) The time of *Euconodontus notchpeakensis* that was a form with the first two-element apparatus.

(IIc) The origin time of *Eoconodontus alisonae* Landing and species with complicated two-element apparatuses from the genus *Cambrooistodus* Miller.

(IId) The appearance time of *Cordylodus primitivus* Bagnoli, Barnes et Stevens and *C. andresi* Viira et Sergeeva, both equipped with first complex (dentate) elements.

At the end of the Batyrbai stage, the Late Cambrian representatives of the *Proconodontus* lineage (*E. alisonae* Landing and species from genera *Proconodontus* Miller and *Cambrooistodus* Miller) became extinct.

(III) The early Tremadocian[1] stage or the *Cordylodus proavus–Cordylodus angulatus* time represents a period of appearance of true cordylodids in the *Proconodontus* lineage. Their successive evolution is traceable

[1] The range of the Tremadocian Stage in terms of Dubinina (1991).

from *C. proavus* Miller to *C. angulatus* Miller. It was accompanied by degradation of proto- and paraconodont assemblages. The conodont evolution at this stage was also marked by the origin of new taxa in the *Hirsutodontus* lineage (beginning from the first occurrence level of *C. proavus*) and in the *Teridontus* phyletic line (at the occurrence level of *C. intermedius*).

Characteristics of the fourth and fifth stages has not been adequately investigated.

(IV) The late Tremadocian or *Drepanoistodus deltifer* stage exemplifies the evolution of conodonts from families Drepanoistodontidae, Oleontodontidae, Acanthodontidae, Westergaardodinidae (genus *Chosonodina* Miller), and from genus *Rossodus* Repetski et Ethington (its family is unknown). The stage also corresponds to the extinction time of paraconodonts.

(V) The early Arenigian or *Paroistodus proteus–Oepikodus evae* stage resulted in appearance and evolution of abundant conodonts represented by 13 families and 22 genera.

The boundary between stages II and III is most distinct and significant. It corresponds to the base of the *Cordylodus proavus* Zone and to the natural Cambrian–Ordovician boundary.

These stages in the evolution of Late Cambrian and Early Ordovician conodonts were used, along with data on distribution of trilobites (Apollonov, 1992), to substantiate the respective stratigraphic subdivisions of the regional and general scales (Apollonov and Chugaeva, 1983; Chugaeva *et al.*, 1989; Apollonov, 1992). The significance of conodonts in stratigraphy is now well approved.

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# Late Albian Radiolarians and Foraminifers from the Mar'ino Section (Outskirts of Simferopol, Crimea): Taxonomic Composition and Stratigraphic Position

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**Abstract**—The first study on taxonomic composition of radiolarians from the upper Albian deposits of Crimea showed that they represent 48 taxa, among which 15 genera and 29 species are classed with spumellarians, and 14 genera and 19 species with nassellarians. The discovered community of radiolarians is termed as an *Orbiculiforma nevadaensis–Dictyomitra konachkendensis* assemblage. It includes species common with those from coeval radiolarian assemblages of the Russian platform, Ukraine, Caucasus, and Sakhalin. The radiolarians in question coexist with planktonic and benthonic foraminifers indicating the late Albian age of their host rocks.

**Key words:** *upper Albian, radiolarians, nassellarians, spumellarians, foraminifers, plankton, benthos*

The presence of radiolarian skeletons in the upper Albian clay deposits of Crimea was noted long ago. Gorbachik was first to roughly estimate their generic composition (see in Drushchits and Gorbachik, 1959). In this work, we present data implying that the late Albian foraminifers coexist in sections of central Crimea with abundant radiolarians representing genera *Rhopalastrum, Hagiastrum, Stylotrochus, Dictyomitra*, and others.

At present, the available data on the taxonomic composition of radiolarians from the region are scarce and imperfect. Bearing in mind that these fossils play a significant role in solving problems of stratigraphy, correlation, and paleoecology, we have made a first attempt to investigate this lacuna.

## BRIEF CHARACTERISTICS OF UPPER ALBIAN IN CRIMEA

Albian deposits are widespread in the piedmont and plain areas of Crimea, where they are penetrated by numerous boreholes. The deposits predominantly consist of clay intercalated with subordinate sandstone beds. Their age is determined on the basis of ammonites, belemnites, bivalves, and foraminifers (Drushchits, 1960; Drushchits *et al.*, 1986). The upper Albian deposits occur within an area extending from the Chernaya River basin of southwestern Crimea to the village of Nanikovo located in the east of the peninsula.

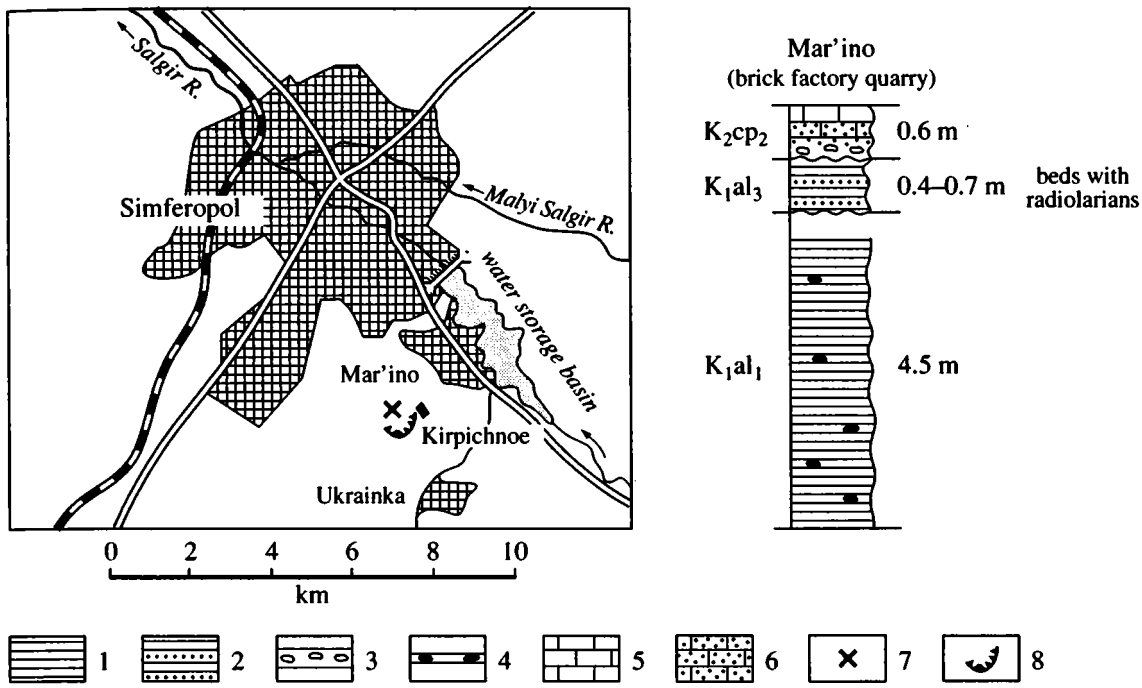
In different piedmont areas of Crimea, the upper Albian deposits vary in thickness from dozens of centimeters (e.g., in the brick factory quarry near the settlement of Mar'ino, central Crimea) to 250 m near the village of Kurskoe in the eastern part of its territory.

Boreholes drilled in the plain areas of Crimea show that the thickness range of upper Albian clay is here 15 to 1300 m or more. In eastern and some areas of central Crimea, these deposits directly overlie the middle Albian clay, but, in other areas, they frequently rest with a hiatus on older sediments.

Despite the scarcity of ammonite fauna, the upper Albian deposits of Crimea are divided into three zonal units (from the base upward): *Hysteroceras orbigny, Pervinqueria inflata*, and *Stoliczkaia dispar* ammonite zones (Drushchits *et al.*, 1986). On the basis of planktonic foraminifers, the lower part of the upper Albian is identified as *Hedbergella infracretacea–H. globigerinellinoides* Beds, and its upper part is attributed to the *Thalmanniella ticinensis* Zone; according to benthic foraminifers, the last unit corresponds to the *Pleurostomella subbotinae–Clavulina gaultina* Zone (Gorbachik, 1986, 1991).

## MATERIALS

The materials used in this work were collected in 1968 and 1971 during the preparation of outcrops and field excursion of the XII European Micropaleontological Colloquium (*Putevoditel' ekskursii*, 1971). Samples of Aptian and Albian clay deposits used to study microfauna were collected in the quarry of the Mar'ino brick factory located in the southeastern sector of suburbs of Simferopol (Fig. 1). All samples yielded abundant shells of foraminifers. In the samples of upper Albian rocks, foraminifers coexist with radiolarians,




Geographic locality and lithology of the upper Albian deposits with radiolarians: (1) calcareous clay; (2) clay with sand interlayers; (3) pebbles; (4) siderite nodules; (5) limestone; (6) sandy limestone; (7) section locality; (8) quarry of the brick factory.

whose abundance is as high as hundreds of skeletons per preparation. The proportions between foraminifers and radiolarians are variable and never equal. One of these groups usually dominates over another in a given sample. This is especially characteristic of clay samples collected near the "Krymneftegazrazvedka" building on Bespalova street of Simferopol in 1968, but not of the Mar'ino section, although rocks from both localities are identical with respect to composition of foraminifer and radiolarian assemblages. The section of the Mar'ino quarry, where the upper Albian clay is 0.4–0.7 m thick, is represented in the studied collection by two samples. Skeletons of radiolarians and foraminifers have been macerated from rock samples by using the routine procedure.

## SECTION OF THE MAR'INO BRICK FACTORY

The Mar'ino brick factory of the Simferopol district (indicated under name "Kirpichnoe" in the figure) utilizes the Aptian and Albian clays as the raw material for production. Its quarry is cut stepwise, and the lower steps expose the middle–upper Aptian clay beds. This interval of the section bearing the abundant planktonic foraminifers is divided into the *Leupoldina protuberans, Globigerinelloides ferreolensis, G. algerianus, Hedbergella trocoidea, Planomalina cheniourensis*, and *Ticinella roberti–P. cheniourensis* zones (Gorbachik, 1986).

Above these zones, there are exposed the gray lower Albian clay beds bearing siderite concretions (figure). The upper Albian clay bed is exposed 4–5 m higher, above the level of landslides and presumable stratiraphic hiatus, because it is of another lithologic type. The bed is composed of black calcareous plastic clay with local sandy admixture. In the upper portion, the bed is green with black spots; locally, it looks variegated when black and green constituents are irregularly distributed. This kind of coloration is characteristic of other localities of upper Albian clay in Crimea. The bed yielded single specimens of planktonic foraminifers *Blefuscuiana* cf. *infracretacea* (Glaessner), *B. globigerinellinoides* (Subbotina), and *Globigerinelloides ultramicrus* (Subbotina). Benthic foraminifers of the bed are predominantly represented by *Pleurostomella obtusa* Reuss, *P. subnodosa* Reuss, and *Lenticulina macrodisca* (Reuss) associated with rare *Gaudryina filiformis* Berthelin and *Glomospirella gaultina* (Berthelin). Analyzing the taxonomic composition of these assemblages, Gorbachik (1986) attributed them to the upper Albian *Hedbergella infracretacea–H. globigerinellinoides* Zone presumably spanning a part or the whole ammonite zone of *Pervinqueria inflata*.

To explain the situation, we should remark that genera *Hedbergella* and *Blefuscuiana* are not synonyms. Banner and Desai (1988) differentiated the latter from the genus *Hedbergella*, because shells of this genus are microporate and have pore canals running through muricae and opening with pores at the surface of the shells.

Plotnikova (1982) studied in detail the Albian deposits of Crimea and their foraminifers. Using a vast

Distribution patterns of late Albian radiolarians in Crimea and some other regions

| Crimea | Russian platform | | Ukraine | Greater Caucasus | Western Sakhalin |
|---|---|---|---|---|---|
| | Moscow syneclise | Volga–Ural anteclise | | | |
| Mar'ino section, the brick factory quarry | Boreholes 9, 105, 107, 195 | Borehole near the village of Kovylkino | Well 1-K, Melitopol district | Konakkhend locality | Naiba River basin |
| | Kazintsova, 1990 | | Gorbunov, 1975 | Aliev, 1965; Abbasov, 1990 | Kazintsova, 1993 |
| *Conosphaera sphaeroconus* Rust | | | | + | |
| *C. haeckeli* Aliev | | + | + | + | |
| *Hexapyramis pantanelli* Squinabol | | | | | |
| *Cromyodruppa concentrica* Lipman | + | | | | |
| *Archaeospongoprunum praelongum* Pessagno | | + | | | + |
| *Ar. cortinaensis* Pessagno | | | | | |
| *Cenodiscus cenomanicus* Aliev | | | | + | |
| *Porodiscus kavilkinensis* Aliev | + | + | | + | + |
| *Euchitonia triradiata* Lipman | | | | | |
| *E. santonica* Lipman | | | | | |
| *Histiastrum membraniferum* Lipman | | | + | | |
| *H. tetracantum* Lipman | | | + | | |
| *Sethocyrtis mosquensis* Smirnova et Aliev | | + | | + | |
| *Diacanthocapsa eugania* Squinabol | + | | | + | |
| *Tricolocapsa kelevudacika* Aliev | | | | + | |
| *T. echitonica* Aliev | | | | + | |
| *Holocryptocanium barbui* Dumitrica | | | | + | + |
| *Dictyomitra ferosia* Aliev | + | + | + | + | |
| *D. momentaria* Aliev | | | | + | |
| *D. konachkendensis* Aliev | | | | + | |
| *Theocampe cylindrica* Smirnova et Aliev | + | + | | + | + |
| *Stichomitra communis* Squinabol | | | | + | |
| *Cyrtocapsa perspicua* Squinabol | | | | + | |
| *C. ambiferia* Aliev | | | | + | |
| *Crolanium cuneatum* (Smirnova et Aliev) | + | + | | + | + |
| *Cr. triangulare* (Aliev) | + | + | | + | + |
| *Amphipyndax stocki* (Campbell et Clark) | | | | | + |
| *Archaeodictyomitra* cf. *apiara* (Rüst) | | | | | |
| *Spongocapsula* (?) *ramoraensis* (Pessagno) | | | | + | + |
| *H. cruciferum* Lipman | | | | | |
| *H. latum* Lipman | | | + | | + |
| *Rhopalastrum trigonale* Lipman | | | | | |
| *R. tricolicum* Aliev | | | | + | |
| *Pentinastrum subbotinae* Lipman | | + | | | |
| *Spongotripus morenoensis* Campbell et Clark | | | + | | |
| *Stylotruchus octacanthus* Lipman | | | | | |
| *Crucella aster* (Lipman) | + | + | + | | |
| *Cr. crux* (Lipman) | + | | | | |
| *Orbiculiforma impressa* (Lipman) | | | | | |
| *Or. maxima* Pessagno | | | | | + |
| *Or. nevadensis* Pessagno | + | + | | | |
| *Or. multangula* Pessagno | + | | | + | + |
| *Or. cachaensis* Pessagno | | | | | |
| *Or. railaensis* Pessagno | + | + | | | |
| *Patulibracchium* cf. *attenuatum* (Lipman) | | | | | |
| *P.* ex gr. *obesum* Pessagno | | | | + | |
| *Cyrtocalpis eurystoma* Rüst | + | | | | |
| *Podocapsa guembeli* Rüst | | | | | |

amount of data on the mountain and plain areas of Crimea, she obtained the representative micropaleontological characteristics of the ammonite *inflata* Zone. According to her conclusion, this zone "... includes an impoverished assemblage of small planktonic forms, among which *Hedbergella infracretacea* (Glaessner), *H. globigerinellinoides* (Subbotina), and *Globigerinelloides ultramicrus* (Subbotina) are dominant. The assemblage characterizes the local foraminiferal zone named after the last species" (Plotnikova, 1982, p. 137).

Younger deposits of the upper Albian with more diverse assemblages of planktonic foraminifers, including the newly appeared *Praeglobotruncana* and *Rotalipora* forms, are missing in the Mar'ino section.

In addition to foraminifers, the upper Albian clay samples yielded the abundant and diverse specimens of radiolarians listed in the Table. The occurrence of late Albian radiolarians in Crimea was noted earlier by Bogaets *et al.* (1980).

In the Mar'ino quarry, the apparent thickness of the upper Albian clay is only 0.4–0.7 m. However, its bed is 4–5 m thick in the road-cut outcrops around Simferopol and about 12 m thick in the excavation at Bespalova street. In all these localities, the clay incorporates similar assemblages of radiolarians and foraminifers.

The eroded surface of Albian deposits is overlain by the light-gray Upper Cretaceous limestone beds incorporating pebbles and glauconite grains at the base. N.I. Maslakova attributed these beds to the upper Campanian (figure). Their apparent thickness is 0.6 m, and they occur below the Paleogene rocks.

## COMPOSITION AND ANALYSIS OF RADIOLARIAN ASSEMBLAGE

The radiolarian assemblage from the upper Albian clay of Crimea is represented by well-preserved fossils. Many of its species have been originally identified in Albian deposits of various regions: 14 in the Russian platform (Aliev and Smirnova, 1969); 14 in the Greater Caucasus (Aliev, 1965); 8 in Italy (Rüst, 1885; Squinabol, 1903); 11 in California (Pessagno, 1977); and one in Romania (Dumitrica, 1970). Dominant in the assemblage are discoid cancellous spumellarians, whose abundance varies from dozens to hundreds of specimens per preparation. The discovered radiolarian community is termed as the *Orbiculiforma nevadaensis–Dictyomitra konachkendensis* assemblage in accordance with names of its index species. The half of the radiolarian assemblage consisting of more than 50 species is represented by spumellarians belonging to 29 species and 15 genera. These, mostly discoid forms have rounded, triangular, tetragonal, and pentagonal skeletons ornamented with, or lacking spines and processes. They represent genera *Porodiscus, Orbiculiforma, Spongotripus, Histiastrum, Crucella, Pentinastrum,* and *Patulibracchium*. Spongy forms and skeletons with patagium are most abundant.

Nassellarins of the assemblage are equally diverse and represented by 19 species of 14 genera. Among them, there are one-chamber (*Cyrtocalpis, Podocapsa*), two-chamber (*Sethocyrtis*), three-chamber (*Diacanthocapsa, Theocampe, Tricolocapsa, Holocryptocanium*), and multichamber skeletons. The latter are exemplified by alveolate (*Dictyomitra, Stichomitra, Amphipyndax*), costate (*Archaeodictyomitra*), tuberculate (*Crolanium*), and cancellate (*Spongocapsula*) forms.

Species known in the Russian platform and Greater Caucasus are most characteristic of the assemblage that looks like a transitional one between the assemblages of these two regions. Late Albian species common for three territories (Russian platform, Crimea, and Greater Caucasus) are *Conosphaera haeckeli* Aliev, *Porodiscus kavilkinensis* Aliev, *Orbiculiforma multangula* Pessagno, *Theocampe cylindrica* Smirn. et Aliev, *Sethocyrtis mosquensis* Smirn. et Aliev, *Diacanthocapsa euganea* Squinabol, *Dictyomitra ferosia* Aliev, *Crolanium cuneatum* (Smirn. et Aliev), and *Cr. triangulare* (Aliev) (Aliev, 1965; Aliev and Smirnova, 1969; Kazintsova, 1990; Kazintsova and Olfer'ev, 1997).

At the same time, the assemblage includes species characteristic of remote regions. For instance, eleven species of our assemblage, such as *Archaeospongoprunum praelongum* Pess., *Porodiscus kavilkinensis* Aliev, *Histiastrum latum* Lipman, *Orbiculiforma maxima* Pess., *Or. multangula* Pess., *Holocryptocanium barbui* Dumitrica, *Theocampe cylindrica* Smirn. et Aliev, *Crolanium cuneatum* (Smirn. et Aliev), *Cr. triangulare* (Aliev), *Amphipyndax stocki* (Campb. et Cl.), and *Spongocapsula (?) zamoraensis* (Pess.) are also present in the late Albian assemblage from the Naiba reference section of western Sakhalin (Zonova *et al.*, 1993).

Five Crimean species (*Archaeospongoprunum praelongum* Pess., *Orbiculiforma multangula* Pess., *O. nevadaensis* Pess., *O. railensis* Pess., and *Patulibracchium obesum* Pess.) also occur in the radiolarian *Kozurium zingulai* Zone of California attributed to the lower–basal upper Albian by Pessagno (1977).

The Crimean assemblage includes species whose late Albian age is approved by coexistence with other fossil groups: with foraminifers, inocerams, and ammonites in the Russian platform (Lipman, 1952; Aliev and Smirnova, 1969; Kazintsova and Olfer'ev, 1997); with foraminifers in the Melitopol district (Gorbunov, 1975); with foraminifers and belemnites in the Greater Caucasus (Aliev, 1965; Azizbekova and Abbasov, 1990); and with ammonites and inocerams in the western Sakhalin (Zonova *et al.*, 1993).

It is also interesting that radiolarian species *Cromyodruppa concentrica* Lipman, *Porodiscus kavilkinensis* Aliev, *Histiastrum latum* Lipman, *Holocryptocanium barbui* Dumitrica, *Dictyomitra ferosia* Aliev, and *Crolanium cuneatum* (Smirn. et Aliev) of the Crimean assemblage are present in the upper Albian

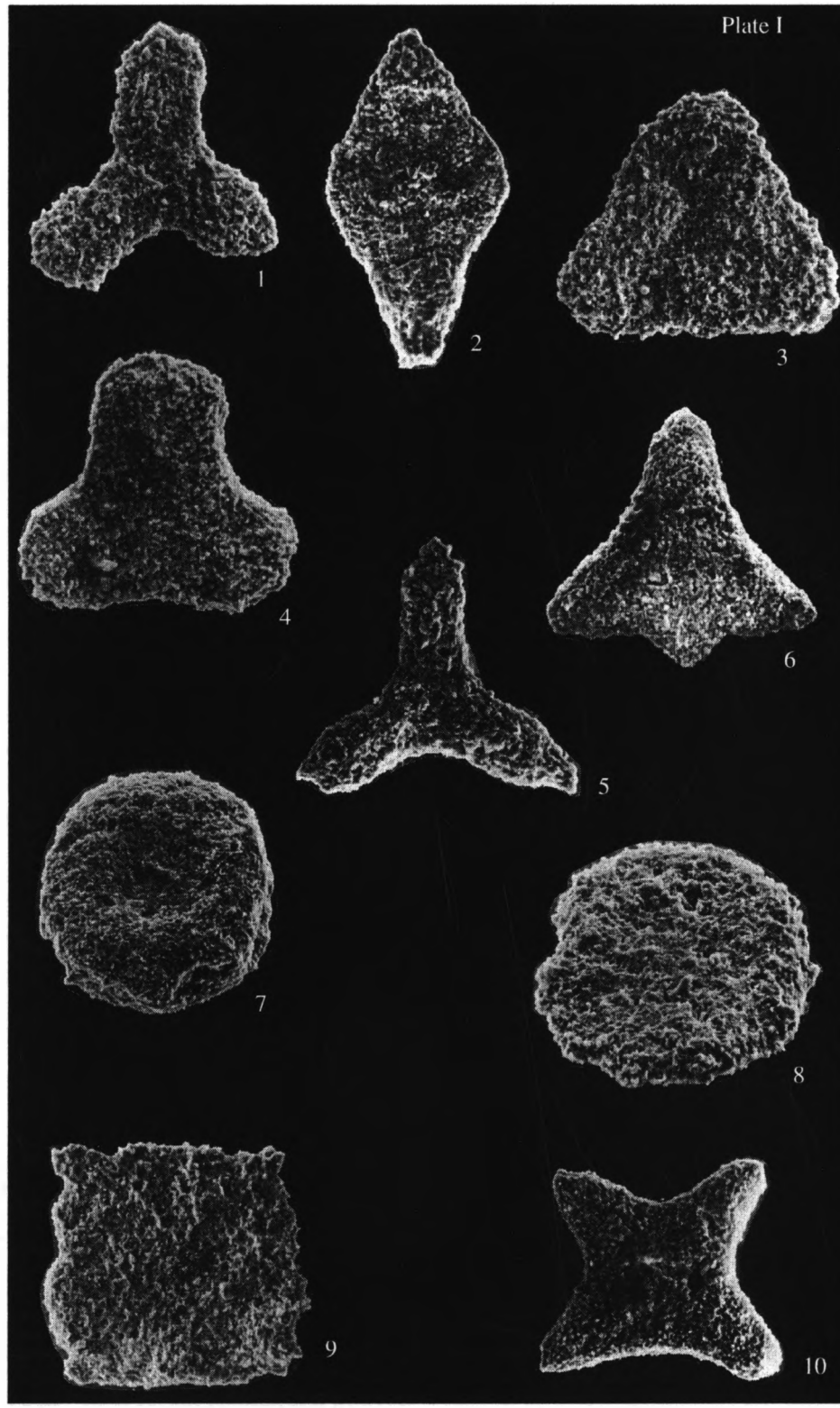



**Plate I.** Late Albian radiolarians from the Simferopol district (Crimea): (1) *Rhopalastrum trigonale* Lipman, ×160; (2) *Histiastrum* cf. *latum* Lipman, ×145; (3) *Spongotripus* cf. *morenoensis* Campbell et Clark, ×125; (4) *Patulibracchium* ex gr. *obesum* Pessagno, ×125; (5) *Patulibracchium* cf. *attenuatum* (Lipman), ×145; (6) *Rhopalastrum* sp., ×145; (7) *Orbiculiforma impressa* (Lipman), ×125; (8) *Orbiculiforma* cf. *nevadaensis* Pessagno, ×145; (9) *Savaryella quadra* (Foreman), ×110; (10) *Crucella aster* (Lipman), ×125.

All species pictured in Plates I and II are from the Mar'ino section located to the southeast of Simferopol, Crimea. The specimens are stored at the Laboratory of Microfauna, All-Russia Research Institute of Geology, collection no. 769.

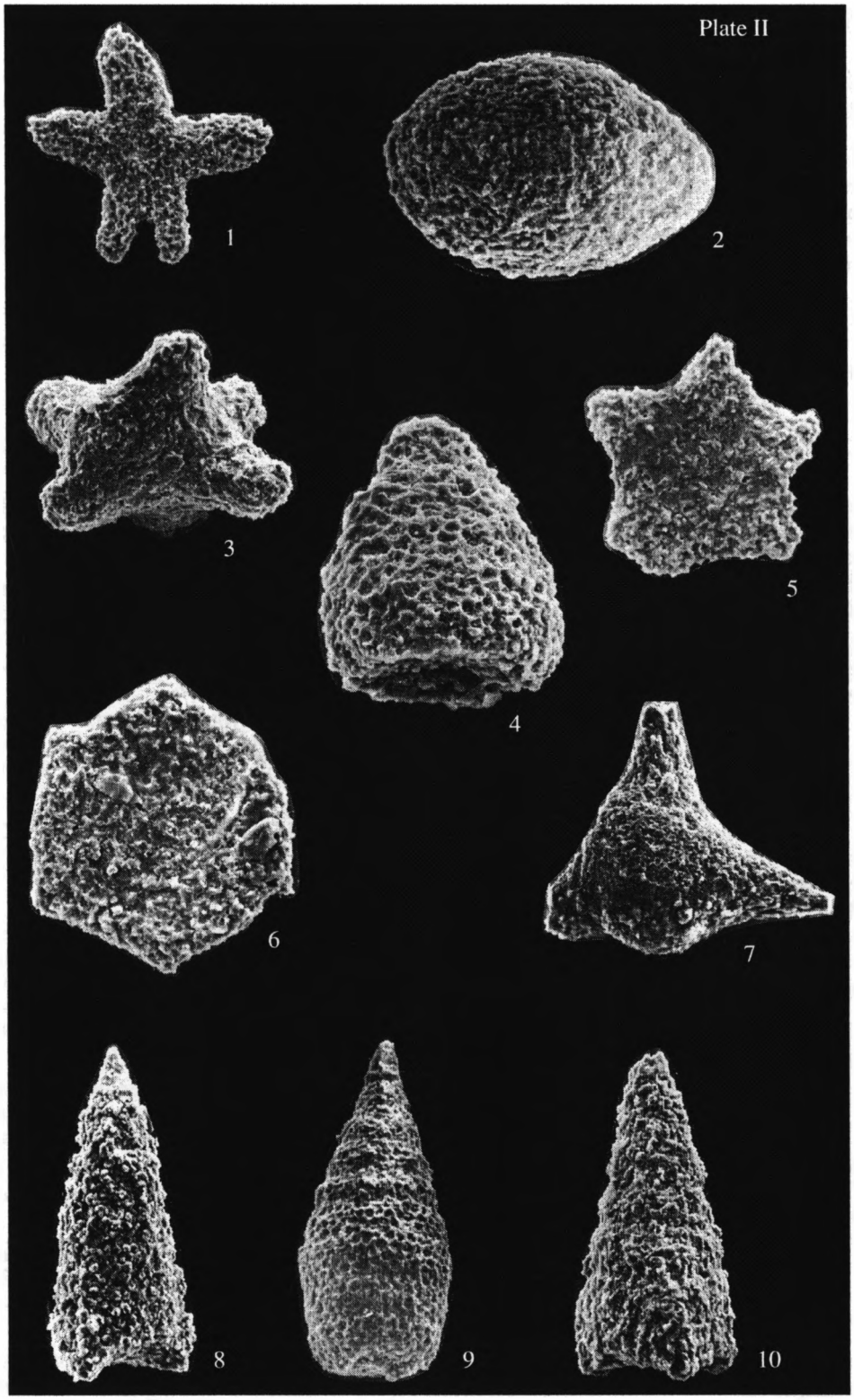



**Plate II.** Late Albian radiolarians from the Simferopol district (Crimea): (1) *Crucella crux* (Lipman), ×430; (2) *Cenodiscus cenomanicus* Aliev, ×180; (3) *Hexapyramis pantanelli* Squinabol, ×110; (4) *Cyrtocalpis eurystoma* Rüst, ×260; (5) *Pentinastrum subbotinae* (Lipman), ×145; (6) *Orbiculiforma multangula* Pessagno, ×180; (7) *Podocapsa quembeli* Rüst, ×180; (8, 10) *Crolanium cuneatum* (Smirnova et Aliev), (8) ×130 and (10) ×145; (9) *Dictyomitra konachkendensis* Aliev, ×125.

Paramonovo Clay of the Moscow district (Alekseev *et al.*, 1996).

Some features of the *Orbiculiforma nevadaensis–Dictyomitra konachkendensis* assemblage, e.g., abundant radiolarian skeletons of compressed morphology (genera *Euchitonia, Histiastrum*, and *Crucella*), suggest that the Crimean sea of the late Albian time was warmer than the coeval seas of the Russian platform, where these forms were extremely rare. Large cupola-shaped nassellarians abundant in our assemblage (genera *Dictyomitra, Stichomitra, Cyrtocapsa*, and *Crolanium*) imply that the Crimean sea was also deeper (nassellarians from the Russian platform are scarce and small). In addition, the Crimean basin appears to be connected with the Mediterranean basins, because eight species of the assemblage are widespread in the Albian deposits of Italy (O'Dogherty, 1995). The outlined similarity between the late Albian radiolarian assemblages from Russian platform, Crimea, Greater Caucasus, and western Sakhalin seems to be the result of past communications between the Boreal and Tethyan seas. This similarity is favorable for interregional stratigraphic correlations.

Most characteristic and representative late Albian radiolarians from Crimea (19 species of 13 genera) are pictured in Plates I and II.

## CONCLUSION

The late Albian radiolarian assemblage from Crimea includes abundant and diverse taxa. Many species of the assemblage show wide geographic, but narrow stratigraphic ranges and have a significant correlation potential. The comparative analysis of radiolarian assemblages from different regions shows that the Crimean basin was connected in the late Albian time with the Boreal and Mediterranean sea basins.

Reviewers V.S. Vishnevskaya and I.A. Basov

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# The Late Golovnino Transgression of the Middle Pleistocene in the Kunashir Island (Kuril Islands)

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**Abstract**—Lithology and facial characteristics of the upper part of the Golovnino Formation in the Kunashir Island (Kuril Islands) are investigated. The distinguished interlayers of volcanic ash may be used as the age and correlation markers. The results of diatom and spore–pollen analyses suggest the middle Pleistocene (Mindel–Riss Interglacial) age of the studied sediments. Two transgressive–regressive cycles of the respective sedimentation period correspond in time to the O-isotope stages 11–9. The elucidated paleogeographic environments indicate that the sea level was 20–25 m higher than now during the first transgressive phase and 5 m higher during the second. The regressive phases are marked by the tuff accumulation and a stratigraphic hiatus. In the middle Pleistocene, the estimated rate of tectonic uprising of the Kunashir Island was 0.28 mm/year. The middle Pleistocene paleogeographic events in the island are correlated with those in Japan, the Chukchi Peninsula, and Alaska.

**Key words:** *middle Pleistocene, diatoms. transgression, paleoclimate, Kuril Islands*

## INTRODUCTION

The problem of glacioeustatic fluctuations of the global sea level is currently of great interest as a factor contributing much to the prognosis of environmental medium evolution. This factor is usually obscured by associated tectonic and eustatic sea-level changes operating within long periods of time. The Kunashir Island with widespread marine deposits of the Pleistocene time, which accumulated during the glacioeustatic transgressions and then were risen up by tectonic movements, is an appropriate object for study of this phenomena.

We investigated upper members of the Golovnino Formation in its stratotype located northward of the synonymous settlement (Zhelubovskii and Pryalukhina, 1964) and in the South Kuril Isthmus, where the terracelike structures are exposed at the ground surface rising about 40 m above the sea (Figs. 1 and 2). The main objective of this work was to identify the age and amplitude of the late Golovnino transgression. The lower part of the formation consist of pumice sands and gravel, whereas the upper one is composed of coarse tuffs and sands. The formation is stratigraphically discordant with respect to the underlying Alekhino Formation. The Golovnino Formation is suggested to be either Pliocene (Piskunov, 1963) or Pleistocene (Pryalukhina, 1961) in age.

## LITHOLOGIC AND FACIAL CHARACTERISTIC OF SEDIMENTS AND TEPHRA COMPOSITION

Sections of the marine deposits in the South Kuril Isthmus (nos. 5800 and 5701) were studied in a quarry located at the foot of the Mt. Otdel'naya, where they are exposed on the terracelike surface at an altitude of about 40 m. A member of tuffaceous diatomite (9.8–13 m) is basal in the section. The refraction index of volcanic glass is 1.504–1.510 in the lower part of the member and 1.507–1.516 in the upper one. The glass corresponds in composition to K-depleted rhyolites ($SiO_2$ = 73.11–74.25%; from here on, concentrations are presented for the water-free residue) of a low-alkalic volcanic series. The content of $K_2O$ (1.38–1.81%) and K/Na ratios (up to 0.61) are higher than in the tephra of the Golovnin Volcano (Fedorchenko *et al.*, 1989). The fine-grained sands are more rich in the heavy fraction (up to 5.6%) than coarse silt (0.02–0.03%). The heavy fraction predominantly consists of magnetite (up to 62%) associated with ortho- and clinopyroxene (up to 12.5 and 19.4%, respectively), amphibole, basaltic hornblende, and apatite.

The middle part of the section (5.0–9.8 m) is composed of tephroid silts incorporating the pumice and sandy-pumice lentils displaying the horizontal wavy lamination. The content of amorphic $SiO_2$ is 13.6–18.96%. Silty rocks (up to 86%) with the average grain size of 0.03–0.05 mm are dominant. The abundance of sandy fraction is not greater than 13% and decreases upward in the section. The degree of material sorting is poor to moderate (Hr = 0.4–0.72), and, along with the low content of heavy fraction (0.1–1.21%), this is indicative of the sediment accumulation under the influence of weak bottom currents. The heavy mineral association is similar in general to that in the tuffaceous diatomites, but the content of ore minerals is a little lower (up to 52.8%). The unit includes several ash

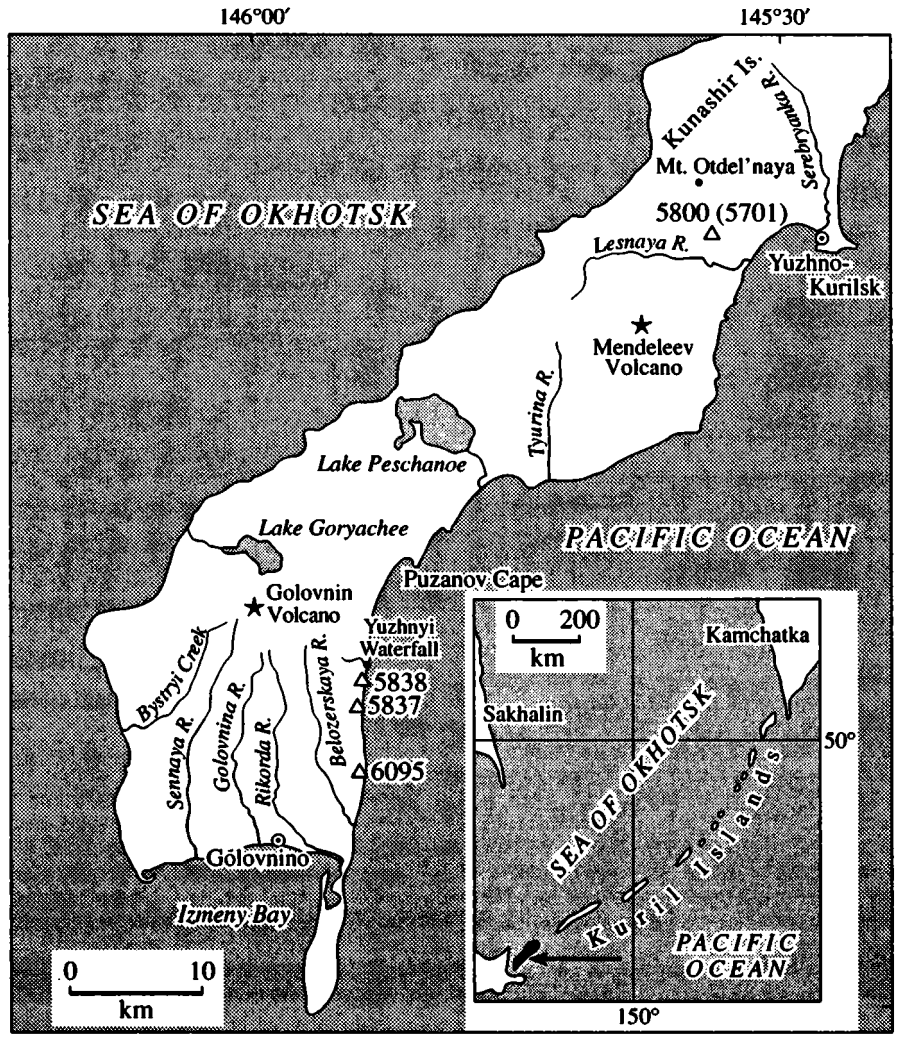


**Fig. 1.** Geographic scheme with indicated sites (triangles) of the Golovnino Formation studied in this work.

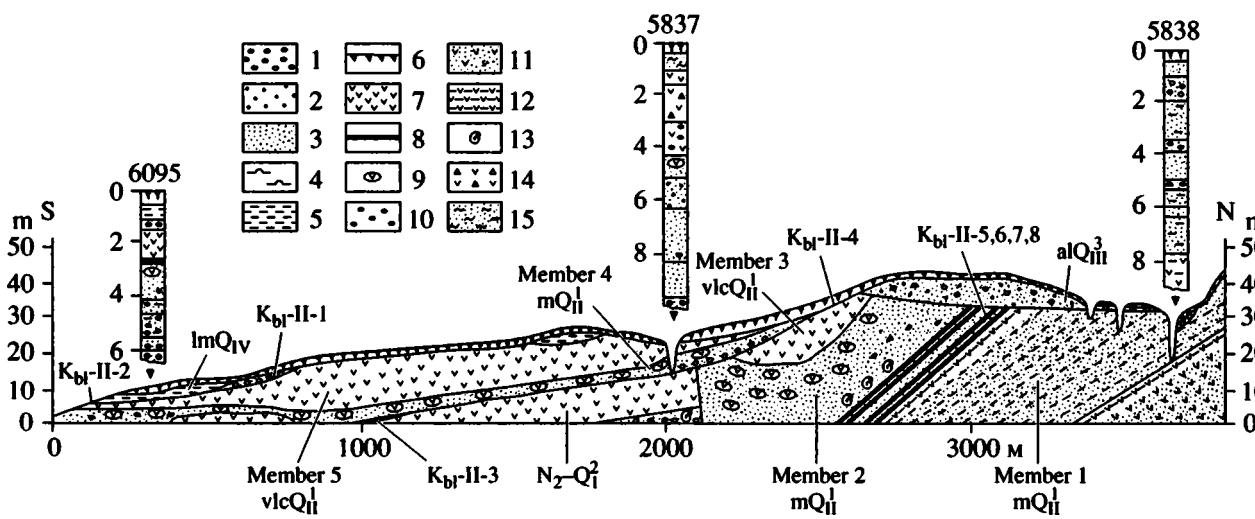


**Fig. 2.** Geological structure of the upper part of the Golovnino Formation southward of the Yuzhnyi Waterfall (altitude 36 m): (1) pebbles and boulders; (2) gravel; (3) sand; (4) tuffaceous diatomite; (5) silt; (6) soil; (7) tuff; (8) ash interlayer; (9) volcanic bombs; (10) pumice; (11) tuffaceous sandstone; (12) tuffaceous silt; (13) fossil mollusks; (14) tuffbreccia; (15) sandy loam.

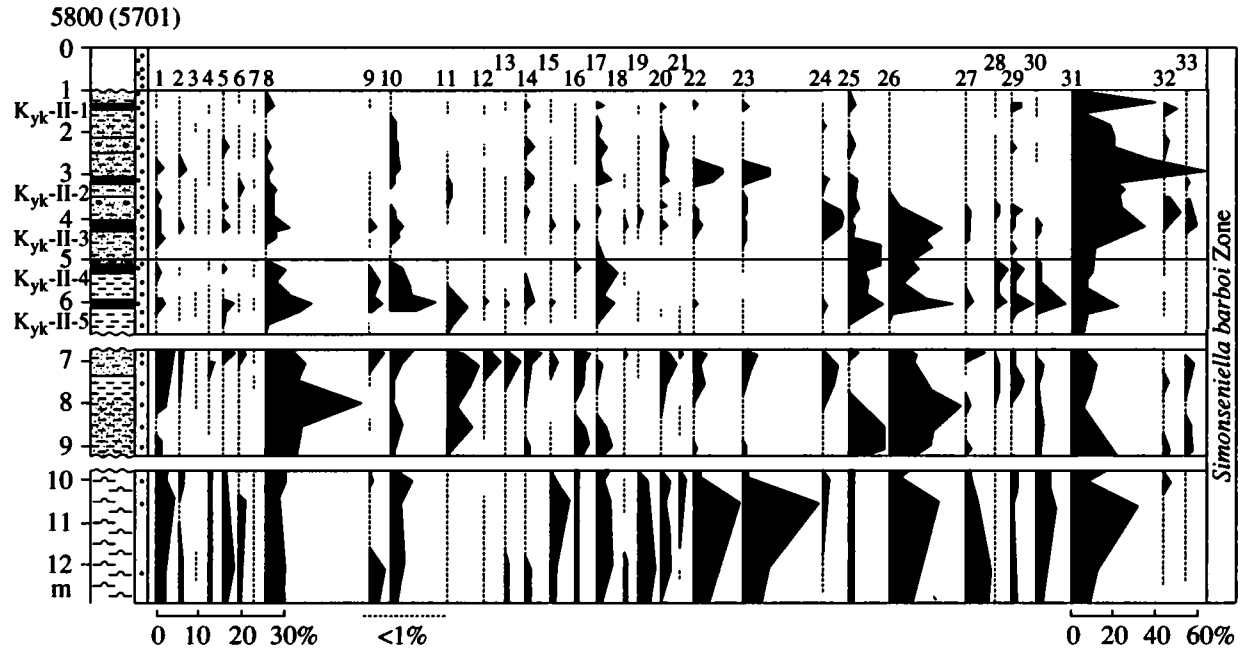


**Fig. 3.** Lithology and distribution of diatoms in the section of Site 5800 (5701), the South Kuril Isthmus (symbols for lithology as in Fig. 2). Group 1 of diatom index species of the *Simonseniella barboi* Zone: (1) *Thalassiosira gravida* Cl. var. *fossilis* Jouse; (2) *Th. nidulus* var. *nidulus* Jouse; (3) *Th. jouseae* Akiba; (4) *Actinocyclus ochotensis* var. *fossilis* Pushkar; (5) *Simonseniella curvirostris* (Jouse) Fenner; (6) *S. barboi* (Brun.) Fenner; (7) *Pyxidicula dimorpha* (Schrader) Pushkar n. comb. (basionim *Stephanopyxis dimorpha* Schrader, Schrader, 1973; diagnosis in Schrader, 1973, p. 711; Holotype, Pl. 15, figs. 9–11, DSDP Leg 18, sample 173-1-4; Paratype: Pl. 15, figs. 19 and 20; Pl. 16, figs. 1–3 and 8–11; Pl. 24, fig. 10). Group 2 of neritic and oceanic arctic- and north-boreal species: (8) *Thalassiosira gravida* Cl. var. *gravida*; (9) *Th. hyalina* (Grun.) Gran; (10) *Th. eccentrica* (Ehr.) Cl., widespread boreal; (11) *Th. kryophila* (Grun.) Jorg.; (12) *Th. antarctica* Comber; (13) *Th. nordenskioeldii* Cl.; (14) *Porosira glacialis* (Grun.) Jorg.; (15) *Coscinodiscus marginatus* Ehr. var. *marginatus*; (16) *Odontella aurita* (Lyngb.) Ag.; (17) *Chaetoceros* sp. sp. (spores); (18) *Neodenticula seminae* (Simonsen) Akiba. Group 3 of neritic and oceanic south-boreal and subtropical species: (19) *Coscinodiscus asteromphalus* Ehr. var. *asteromphalus*; (20) *C. radiatus* Ehr. + *C. perforatus* Ehr.; (21) *Azpeitia nodulifera* (A. Sm.) Fryxel et Sims; (22) *Actinocyclus curvatulus* Janish var. *curvatulus*, widespread boreal; (23) *A. divisus* (Grun.) Hust.; (24) *Actinoptychus undulatus* (Bailey? Kuetz.) Ralfs in Pritchard. Group 4 of subtidal benthic species of arctic- and north-boreal types: (25) *Paralia sulcata* (Ehr.) Cl. var. *sulcata*; (26) *Delphineis kippae* Sancetta; (27) *Grammatophora oceanica* (Ehr.) Grun. var. *oceanica*; (28) *Trachyneis aspera* Cl. var. *aspera*; (29) *Diploneis smithii* (Breb.) Cl. var. *smithii*; (30) *Cocconeis scutellum* Ehr. var. *scutellum*. Group 5 of subtidal benthic south-boreal species: (31) *Hyalodiscus obsoletus* Shesh.-Por.; (32) *Arachnoidiscus ehrenbergii* Ralfs; (33) *Navicula marina* Ralfs + *Lyrella lyra* (Ehr.) Karaeva var. *lyra* (1% = 3 diatom valves).

interlayers that can be used as the age and correlation markers (Fig. 3).

From the base upward, we distinguished the following reference levels.

*Ash Bed $K_{yk}$-II-5* (3–4 cm thick): yellow-gray silt of the rhyolitic composition ($SiO_2$ = 71.45%; $K_2O$ = 1.79%). The refraction index of volcanic glass is 1.507–1.516. The low content of heavy fraction (0.05–0.17%) indicates that the ash source was located far away. Magnetite (71.1%) dominant in this fraction and rather abundant hornblende (15.6%) associate with pyroxenes and apatite.

*Ash Bed $K_{yk}$-II-4* (4 cm thick): light-gray silt of the rhyolitic composition ($SiO_2$ = 73.67%). The average grain size is 0.02 mm. The refraction index of volcanic glass is 1.504–1.507. Heavy fraction consisting of magnetite (52.8%), hornblende, and ortho- and clinopyroxene (1.8% in total) represents 0.22% of the rock volume. The content of amorphic $SiO_2$ is 8.08%.

It seems that material of the beds was derived from volcanoes of the Hokkaido Island. They were quite active in the middle Pleistocene, and their tephra is enriched in $K_2O$ and amphiboles (Okumura, 1988).

The upper part of the section (1–5.0 m) consists of silt intercalated with fine-grained sands bearing pumice and ash interlayers. The average grain size varies here from 0.04 to 0.13 mm, and the sandy fraction content can be as high as 50%. The material is poorly sorted (Hr = 0.52–0.83). It was presumably accumulated in a shallow shelf at a lesser depth than the underlying deposits. This is also evident from a higher content of its heavy fraction (up to 4%). The sedimentation rate was high, and volcanoes of the Kunashir Island were the likely source structures of the pyroclastic material. The heavy fraction includes clino- (up to 37.8%) and orthopyroxene (up to 37.1 %), magnetite (up to 33.9%), and amphibole (up to 9.7%). The content of amorphic $SiO_2$ is 4.8–8.96%. The upper part of the section includes several persistent ash interlayers.

*Ash Bed $K_{vk}$-II-3* (30 cm thick): predominantly fine silt (up to 84.6%) of the dacitic composition ($SiO_2$ = 68.63%). The refraction index of volcanic glass is 1.508–1.511. The heavy fraction is enriched in clinopyroxene (41.2%) and amphibole (9.3%). The content of amorphic $SiO_2$ is 6.8%.

*Ash beds $K_{vk}$-II-2* (1–2 cm thick) together with the upper pumice bed are dacitic in composition ($SiO_2$ = 66.11–66.37%), and the $K_2O$ content in all these beds is low (0.86–1.38%). The refraction index of volcanic glass is 1.509–1.512 in the lower ash bed and 1.510–1.512 in the upper one. Their heavy mineral association is similar to that in the enclosing rocks, and this suggests a common source of tephra.

The marine deposits are overlain by the yellow-brown sandy loam and soil (1–0 m) incorporating lenses of redeposited tephroid material.

In the southeastern coast of the island, within an area extending from the Yuzhnyi Waterfall (altidude 36 m) to the Belozerskaya River mouth, we studied the upper part of the Golvnino Formation stratotype. Members of the section (Fig. 2) discriminated here are decribed below.

*Member 1* (about 45 m thick) concordantly overlies the gray-green marine tuffaceous sandstone of the middle part of the Golovnino Formation. The member consists of tephroid silt intercalated with sands of variable grain size bearing well rounded pumice pebbles and gravel. Its topmost bed (up to 2 m thick) is represented by tuffaceous silt with lenses of tuffbreccia. In the lower part of the member at Site 5838, the greenish gray tuffaceous siltstone (Fig. 2) with a low content of heavy fraction (0.01–0.14%) rich in ore minerals (68.9%) is exposed. This bed is overlain by tuffaceous silt and sand beds (grain-size range 0.08–0.19 mm) with poorly sorted material (Hr = 0.75–0.92). These rocks accumulated under conditions of a low hydrodynamic activity in a shelf located near the foot of the Golovnin Volcano (Markhinin, 1957). Tephroid material of the rock is dacitic in composition ($SiO_2$ = 70.43–75.61%). The content of amorphic $SiO_2$ is 6.08–10%. The heavy mineral association is similar to that in deposits of Site 5800 (5701).

*Member 2* (about 25 m thick) conformably rests on the previous one and consists of greenish yellow loose sandstone with lenticular bedding and black laminae of heavy minerals in its basal part; above this level the bedding is of the horizontal wavy type. The bed yielded shells of marine mollusks. The lower part of the member includes beds with pumice inclusions and volcanic bombs. The beds are intercalated with four interlayers ($K_{bl}$-II-5)–($K_{bl}$-II-8) of rhyolite volcanic ash ($SiO_2$ = 70.66–73.73%) as thick as 5–7 cm. The uppermost interlayer is more rich in $K_2O$ (3.52%; K/Na = 1.15) than others and can be related in origin to one of volcanic eruptions in Hokkaido. Pumice from the member is close in composition to tephra of the Golovnin Volcano (Fedorchenko *et al.*, 1989). Material of these deposits is well sorted, and they appear to be accumulated at a depth no greater than 30 m in a period of relatively weak volcanic activity, when the sea managed to rework the volcanic material. In their characteristics, the deposits are similar to recent sands of the South Kuril Strait (Murdmaa, 1961).

The upper part of the member was accumulated during a new cycle of volcanic activity, as it is evident from an increased amount of pumice and angular volcanic bombs (15–20 cm large), presumably produced by destruction of the crater plug and some elements of the volcanic edifice. The deposits are vaguely laminated because of a high sedimentation rate. The Golovnin Volcano, whose cone rose at that time above the sea level, was the main source of sedimentary material.

*Member 3* (about 16 m thick) concordantly overlies Member 2. Tephroids from its lower part consist of well rounded pumice pebbles and gravel ($SiO_2$ = 66.96%, $K_2O$ = 0.58%). The upper part consists of pumice tuff ($K_{bl}$-II-4). Gravel beds and contact zones of the member are ferruginate, and the member seems to have been deposited during an immense eruption of the Golovnin Volcano. Its coarse-grained tephroids accumulated in a coastal zone, and tuffs, in subaerial settings.

*Member 4* (up to 6 m thick) discordantly overlies members *2* and *3*, presumably together with the Pliocene deposits of volcanogenic–sedimentary origin (Fig. 2). At Site 5837, the coastal shingle of the member is overlain by yellowish gray tephroid sands with a distinct horizon of volcanic bombs. The grain size in the sands varies between 0.14 and 0.4 mm and is minimal in the middle part of the member. The material is poorly sorted (Hr = 0.77–0.85), especially in the bomb horizon (Hr = 0.89–0.94), and suggests a close proximity of the site to the source of tephra (Golovnin Volcano) and an intense influx of tephra in a zone of marine settings, where the rate of sedimentation was high. The deposits are rich in heavy minerals (up to 25%). The peak content of this fraction (48.31%) is recorded in the matrix of coastal shingle, where it consists mainly of magnetite (75.7%). In the coarse silty fraction of sands, magnetite (59.8%) occurs in association with orthopyroxene almost equally abundant (up to 52.2%). Other heavy minerals are hornblende, ilmenite, hematite, and apatite. The content of amorphic $SiO_2$ is 0.24–6.8%.

Along the strike of the member beds, the described sands grade into the silty sands of a deeper facies (Site 6095). The latter rest on the basal bed $K_{bl}$-II-3 of the light-gray ash (thickness up to 0.5 m). The topmost bed $K_{bl}$-II-2 consists of white ash (silty sand) incorporating lapilli. The percentage of heavy fraction in the ash grows upward from 0.11 to 1.95%. The fraction is rich in ortho- and clinopyroxene, magnetite, and apatite, the latter representing up to 4.1% of its volume.

*Member 5* (up to 20 m thick) concordantly overlying Member *4* includes the basal bed of pumice tuff and tuffbreccia resting on the latter. The tuff of dacite–rhy-

olite composition ($SiO_2$ = 67.5–71.99%) shows a low content of $K_2O$ (0.49–0.66%). The rock and bombs of Member *4* are identical in chemical composition and were formed apparently during the same eruptive cycle. The refraction index of volcanic glass is 1.513–1.520. The heavy fraction not abundant in the rocks is mainly concentrated in fine-grained sands (1.03–3.23%). Orthopyroxene (up to 64.8%) and magnetite (up to 27.5%) predominant in the fraction associate with less abundant clinopyroxene (1.4–2.1%). The lower part of the member accumulated in a shallow marine setting, and the upper one is of subaerial genesis. These deposits recorded a phase of the Golovnin Volcano activity, when the sea level droped. Together with rocks of Member *4*, they fill in an overdeepened valley.

## DIATOM ASSEMBLAGES

The diatom analysis was performed for 68 rock samples from the Golovnino Formation.

*Section of the South Kuril Isthmus (sites 5701 and 5800).* The studied 40 samples from the sites yielded 228 species and varieties of 64 diatom genera. The percentage of marine forms amid them is 90–95%. Genera *Thalassiosira, Coscinodiscus, Hyaloduscus, Paralia, Actinocyclus*, and *Delphineis* are dominant in all diatom assemblages. Sporadic redeposited valves of Neogene diatoms represent the following taxa: *Coscinodiscus marginatus* Ehr. var. *fossilis* Jouse, *C. pustulatus* Mann, *Cosmiodiscus insignis* Jouse, *C. intersectus* Jouse, *Pyxidicula zabelinae* (Jouse) Makar. et Moiss, *P. schenckii* (Kanaya) Streln. et Nikol., *Pseudopyxilla americana* (Ehr.) Forti, and *Stictodiscus kittonianus* Gran.

Chryzophyta from the isthmus area are represented by silicoflagellates *Paradictyocha polyactus* (Ehr.) Freng. f. *completa* Freng., *Distephanus speculum* (Ehr.) Haeckel var. *speculum*, *D. speculum* var. *septenarius* (Ehr.) Jorg., *D. octonarius* (Ehr.) Defl. var. *polyactus* (Jorg.) Gleser, and *Dictyocha fibula* (Ehr.) var. *fibula*. Their total abundance is as high as 5–7%.

Four diatom assemblages succeeding upward one another are distinguished in the section (Fig. 3).

*Assemblage I* (13–9.8 m) is characterized by presence of extinct species (Koizumi, 1986; Barron, 1992) *Thalassiosira gravida* var. *fossilis* (6.7%), *Th. nidulus* var. *nidulus* (1.3%), *Th. jouseae* (singular), *Actinocyclus ochotensis* var. *fossilis* (1.3%), *Simonseniella (Rhizosolenia) curvirostris* (2.7%), *S. barboi* (1.7%), and *Pyxidicula dimorpha* (Schrader). Planktonic boreal species *Actinocyclus curvatilus* var. *curvatilus* (10.7%) and *A. divisus* (19.3%) are dominant. The diatom group of subtidal zone is mostly represented by the north-boreal *Delphineis kippae* (12%) and south-boreal *Hyalodiscus obsoletus* (33%).

*Assemblage II* (9.8–6.7 m) includes the same extinct species (up to 6.7%). Its dominant forms are the arctic- and north-boreal *Thalassiosira gravida* var. *gravida* (24%) and *Th. kryophila* (7.7%), the north-boreal subtidal *Paralia sulcata* var. *sulcata* (9%) and *Delphineis kippae* (17%), and also the south-boreal *Hyalodiscus absoletus* (20.6%).

*Assemblage III* (6.7–5.0 m). Amid the extinct species of the assemblage, only *Thalassiosira gravida* var. *fossilis* and *Simonseniella curvirostris* remain at the abundance level of 2–3%, whereas others are sporadic. The neritic arctic- and north-boreal *Th. gravida* var. *gravida* (11.7%), *Th. cryophila* (5%), and *Chaetoceros* spp. (spores, 6.3%) together with boreal-tropical-notal *Th. eccentrica* (12%) represent most typical forms of the assemblage. The subtidal group includes the north-boreal species *Paralia sulcata* var. *sulcata* (8.3%) and *Delphineis kippae* (15.3%). The abundance of *Hyalodiscus obsoletus* is at the same level as before (21.6%). The absolute abundance of diatoms grows in the ash interlayers up to 10–11 millions of valves per 1 g of sediment (Fig. 6)

*Assemblage IV* (5.0–1.0 m). The occurrence frequency of species extinct in the Pleistocene (Barron, 1992; Barron and Gladenkov, 1995) is below 2%. In the lower part of the member (5.0–3.5 m), *Delphineis kippae* (12.7%) and *Paralia sulcata* (8%) are dominant forms of the subtidal cold-water group. The subtidal boreal group is represented by *Hyalodiscus obsoletus* (36%), and *Actinoptychus undulatus* (5%) is the characteristic form of neritic taxa. Oceanic boreal species *Actinocyclus curvatulus* (7.3%) and *A. divisus* (8%) occur in the middle part of the member (3.5–2.8 m). *Hyalodiscus obsoletus* (up to 60%) is dominant in the interval of 2.8–1.0 m.

The upper part of the section (1.0–0 m) yielded a mixed assemblage of redeposited marine species from genera *Coscinodiscus* and *Thalassiosira* associated with the fresh-water halophobes representing genera *Pinnularia, Eunotia*, and *Cymbella*.

## CLIFF SECTIONS OF THE SOUTHEASTERN COAST (SITES 5838, 5837, AND 6095)

Sixteen samples collected at Site 5838 yielded 87 species and varieties of 38 Bacillariophyta genera. Marine diatoms represent 95.7% of them (Fig. 4) and are divided into the three assemblages described below from the secton base upward.

*Assemblage I* (9.1–7.8 m). *Thalassiosira gravida* var. *gravida* (14%) and *Bacterosira fragilis* (10%) are dominant in the assemblage. Diatoms of the subtidal zone are represented by the north-boreal forms *Paralia sulcata* (5.5%), *Delphineis kippae* (6.5%), and *Diploneis interrupta* + *subcincta* (6%). Other characteristic species are *Pyxidicula dimorpha* (sporadic), *Thalassiosira gravida* var. *fossilis* (4%), *Th. nidulus* var. *nidulus* (2%), *Actinocyclus ochotensis* var. *fossilis* (occasional), *Simonseniella curvirostris* (4%), and *S. barboi* (1.5%).

*Assemblage II* (7.8–5.0 m) is very impoverished and includes only rare specimens (1–2%) of *Thalassiosira*

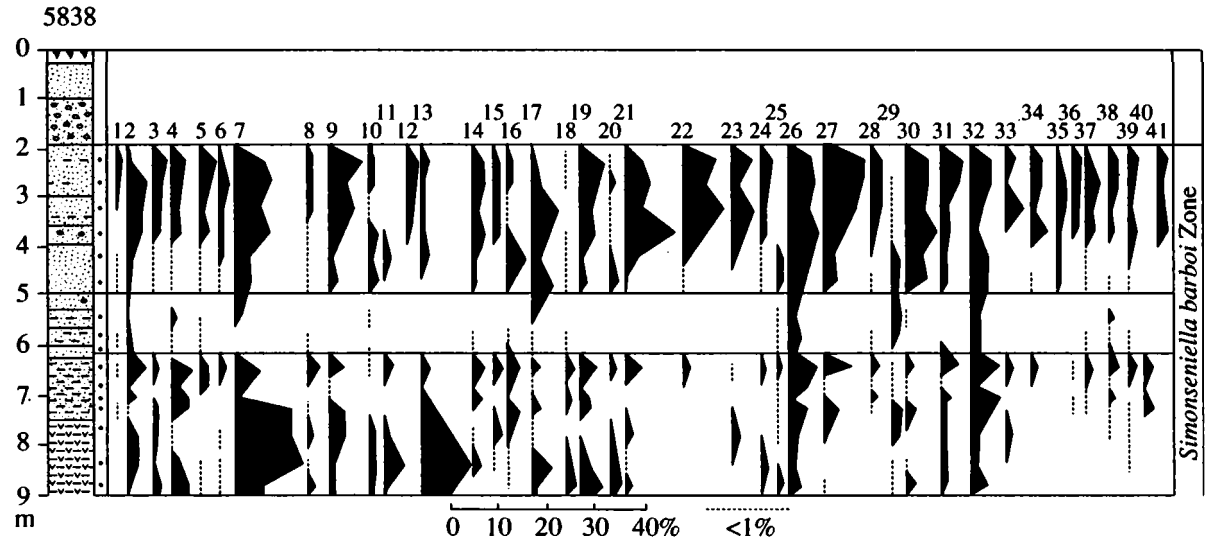


**Fig. 4.** Lithology and distribution of diatoms in the section of Site 5838, the Golovnin Cliff of the Yuzhnyi Waterfall (symbols for lithology as in Fig. 2). Group 1 of diatom index species of the *Simonseniella barboi* Zone: (1) *Pyxidicula dimorpha* (Schrader) Pushkar n. comb.; (2) *Thalassiosira gravida* Cl. var. *fossilis* Jouse; (3) *Th. nidulus* var. *nidulus* Jouse; (4) *Simonseniella curvirostris* (Jouse) Fenner; (5) *S. barboi* (Brun.) Fenner; (6) *Actinocyclus ochotensis* var. *fossilis* Pushkar. Group 2 of neritic and oceanic arctic- and north-boreal species: (7) *Thalassiosira gravida* Cl. var. *gravida*, (8) *Th. hyalina* (Grun.) Gran; (9) *Th. eccentrica* (Ehr.) Cl., widespread boreal and notal; (10) *Th. kryophila* (Grun.) Jorg., (11) *Th. antarctica* Comber; (12) *Th. nordenskioeldii* Cl.; (13) *Bacterosira fragilis* Gran; (14) *Coscinodiscus marginatus* Ehr. var. *marginatus*; (15) *Coscinodiscus oculus-iridis* Ehr.; (16) *Chaetoceros* sp. sp. (spores); (17) *Rhizosolenia hebetata* f. *hiemalis* Gran; (18) *Thalassiotrix longissima* Cl. et Grun.; (19) *Neodenticula seminae* (Simonsen) Akiba. Group 3 of neritic and oceanic south-boreal and subtropical species: (20) *Thalassiosira oestrupii* (Ostf.) Hasle; (21) *Coscinodiscus asteromphalus* Ehr. var. *asteromphalus*; (22) *C. radiatus* Ehr. + *C. perforatus* Ehr.; (23) *Actinocyclus divisus* (Grun.) Hust.; (24) *Thalassionema nitzschioides* Grun.; (25) *Rhizosolenia styliformis* Bright. Group 4 of subtidal benthic species of arctic- and north-boreal types: (26) *Paralia sulcata* (Ehr.) Cl. var. *sulcata*; (27) *Delphineis kippae* Sancetta; (28) *Grammatophora oceanica* (Ehr.) Grun. var. *oceanica*; (29) *Navicula distans* W. Sm., (30) *Trachyneis aspera* Cl. var. *aspera*; (31) *Diploneis smithii* (Breb.) Cl. var. *smithii*; (32) *D. smithii* (Breb.) Cl. var. *smithii* + *D. interrupta* (Kuetz.) Cl. var. *interrupta*. Group 5 of subtidal benthic south-boreal species: (33) *Cyclotella striata* (Kuetz.) Grun var. *striata*; (34) *Actinoptychus undulatus* (Bailey? Kuetz.) Ralfs in Pritchard (35) *Aulacodiscus affinis* Grun.; (36) *Navicula marina* Ralfs; (37) *Lyrella hennedyi* (W. Sm.) Kar. var. *hennedyi*; (38) *Diploneis crabro* Ehr. var. *crabro*; (39) *D. weissflogii* (A. Sm.) Cl.; (40) *Achnanthes brevipes* var. *intermedia* (Kuetz.) Cl.; (41) *Nitzschia angularis* W. Sm. (1% = 2 diatom valves).

*gravida*, *Rhizosolenia hebetata* f. *heimalis*, *Simonseniella curvirostris*, *Paralia sulcata*, *Navicula distans*, *Diploneis subcincta*, and *D. interrupta*.

*Assemblage III* (5.0–2.0 m) consists of 67 species and their varieties. Planktonic stenohaline diatoms are most abundant (up to 65%). Dominant among the latter are the arctic- and north-boreal *Thalassiosira gravida* var. *gravida* (7.5%) and *Neodenticula seminae* (5%); *Rhizosolenia hebetata* f. *hiemalis* (5.5%), the boreal-tropical-notal *Th. eccentrica* (7%); and south-boreal to subtropical taxa *Coscinodiscus asteromphalus* var. *asteromphalus* (12%), *C. radiatus* (5.5%), and *C. perforatus* (4%). The north-boreal euryhaline species *Paralia sulcata* (5.5%), *Delphineis kippae* (8.5%), and *Trachyneis aspera* (7%) together with south-boreal *Actinoptychus undulatus* (4%) and *Cyclotella striata* (3.5%) are dominant in the subtidal flora.

Twelve samples from Site 5837 show that the extinct diatom species are sporadic in its section (Fig. 5), and only one of them, *Simonseniella barboi*, constitutes up to 2% of the assemblage. Its absolute dominant (interval of 9.5–6.6 m) is the north-boreal subtidal species *Paralia sulcata* var. *biseriata* (86.8%) that seems to become extinct in the middle–initial early Pleistocene. Neritic *Pyxidicula turris* (3%), *P. nipponica* (2%), *Thalassiosira gravida* var. *gravida* (3%), and oceanic *Coscinodiscus marginatus* (2.5%) and *Actinocyclus divisus* (3%) are other noticeable components of the assemblage also incorporating the redeposited Neogene forms of *Pyxidicula* and *Coscinodiscus* genera (up to 10% in sum). At the top of this section, diatoms are scarce. The assemblage represents a community of a shallow coastal zone near the river mouth, as it is evident from presence of rheophilic species from genera *Pinnularia*, *Navicula*, and *Meridion*.

At Site 6095, we distinguished two diatom assemblages (Fig. 2).

*Assemblage I* (3.45–2.85 m) consists of 188 diatom species and varieties. A rather high occurrence frequency is characteristic in this case of two *Simonseniella* forms: *S. curvirostris* (12.5%) and *S. barboi* (7.5%); the extinction time of both was in the middle Pleistocene. Other extinct taxa are *Thalassiosira gravida* var. *fossilis* (11.2%), *Th. nidulus* var. *nidulus* (2%), *Actinocyclus ochotensis* var. *fossilis* (3%), and *Pyxidicula dimorpha* (3%). *Thalassiosira gravida*

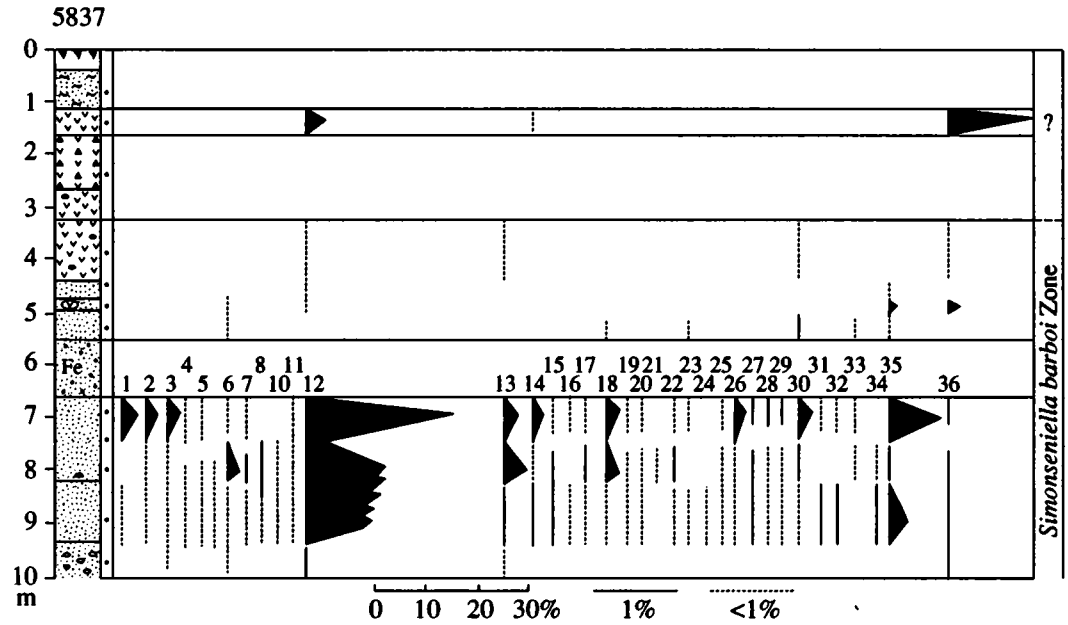


**Fig. 5.** Lithology and distribution of diatoms in the section of Site 5837, the Golovnin Cliff (symbols for lithology as in Fig. 2). Diatoms: (1) *Pyxidicula turris* (Grev et Angst.) Str. et Nik. var. *turris*; (2) *P. nipponica* (Gran et Yendo) Str. et Nik.; (3) *Thalassiosira gravida* Cl. var. *gravida*; (4) *Th. gravida* Cl. var. *fossilis* Jouse; (5) *Th. nidulus* var. *nidulus* Jouse; (6) *Th. jouseae* Akiba; (7) *Th. eccentrica* (Ehr.) Cl.; (8) *Th. pacifica* Gran et Angst; (9) *Th. latimarginata* Makar.; (10) *Porosira glacialis* (Grun.) Jorg.; (11) *Bacterosira fragilis* Gran; (12) *Paralia sulcata* var. *biseriata*; (13) *Hyalodiscus obsoletus* Shesh.-Por.; (14) *Coscinodiscus marginatus* Ehr. var. *marginatus*; (15) *Coscinodiscus radiatus* Ehr.; (16) *Actinocyclus ochotensis* var. *fossilis* Pushkar; (17) *A. curvatulus* Janish var. *curvatulus*; (18) *A. divisus* (Grun.) Hust.; (19) *Arachnoidiscus ehrenbergii* Ralfs; (20) *Actinoptychus undulatus* (Bailey? Kuetz.) Ralfs in Pritchard; (21) *A. undulatus* var. *tamanica* Jouse; (22) *Arachnoidiscus ehrenbergii* Ralfs; (23) *Odontella aurita* (Lyngb.) Ag.; (24) *Chaetoceros compressus* Laud.; (25) *Simonseniella curvirostris* (Jouse) Fenner; (26) *S. barboi* (Brun.) Fenner; (27) *Rhaphoneis amphiceros* Ehr. var. *amphiceros*; (28) *Delphineis kippae* Sancetta; (29) *Rhabdonema arcuatum* Kuetz. var. *arcuatum*; (30) *Grammatophora oceanica* (Ehr.) Grun. var. *oceanica*; (31) *Lyrella lyra* (Ehr.) Kar. var. *lyra*; (32) *Trachyneis aspera* Cl. var. *aspera*; (33) *Cocconeis costata* Greg.; (34) *C. vitrea* Brun.; (35) redeposited Neogene marine species; (36) fresh-water diatoms (1% = 3.5 diatom valves).

(26.7%), *Th. eccentrica* (9.8%), and *Actinocyclus divisus* + *curvatulus* (16%) are dominant taxa of the assemblage. The abundance of subtidal diatoms is below 10%. Tropical diatoms *Azpeitia nodulifera* and *Thalassiosira lineata* (2–3% each), together with south-boreal and subtropical *Coscinodiscus asteromphalus* (up to 3%) and *Th. oestrupii* (2%), are also significant components of the assemblage.

*Assemblage II* (2.85–0.7 m) is not very diverse. It includes 37 species, whose occurrence frequency is low. Noticeable forms of the assemblage are *Pyxidicula dimorpha, Simonseniella barboi, S. curvirostris*, and *Actinocyclus ochotensis*. Sporadic specimens represent the following taxa: moderately thermophilic *Coscinodiscus asteromphalus* and *Thalassiosira pacifica*; widespread boreal *Actinocyclus curvatulus, Actinoptychus undulatus*, and *Thalassiosira eccentrica*; and cryophilic *Coscinodiscus marginatus, Th. gravida* + *gravida* var. *fossilis*, and *Porosira glacialis*. The group of rare subtidal diatoms consists of thermophilic *Lyrella lyra, Navicula marina, Coscinodiscus nitidus, C. granulosus*, and *Auliscus caelatus* associated with cryophilic *paralia sulcata* var. *sulcata, P. sulcata* var. *biseriata, Diploneis smithii*, and *D. interrupta*.

Thus, the Pleistocene diatom succession of the Kunashir Island is divisible into four assemblages. All or some of these assemblages are recognizable in all sections studied.

## ECOLOGIC TYPES AND CLIMATIC SERIES OF DIATOMS

Identifying the ecologic tolerance of different species groups from our diatom assemblages, we used information about the ecologic parameters of these fossils (Jouse, 1962; Koizumi, 1986) and distribution patterns of diatoms in the present-day coenoses and in the Pleistocene sediments of the North Pacific (Jouse, 1962; Sancetta, 1982). Their tolerance to the temperature was evaluated on the basis of the coefficient suggested by Kanaya and Koizumi (1966):

$$Td = 100Xw/(Xw + Xc),$$

where $Xw$ is the amount of valves of thermophilic and moderately thermophilic species of an assemblage, and $Xc$ is the same parameter for the cryophic forms.

The $Xw$ group includes *Thalassiosira oestrupii, Th. lineata, Th. pacifica, Coscinodiscus asteromphalus* var.

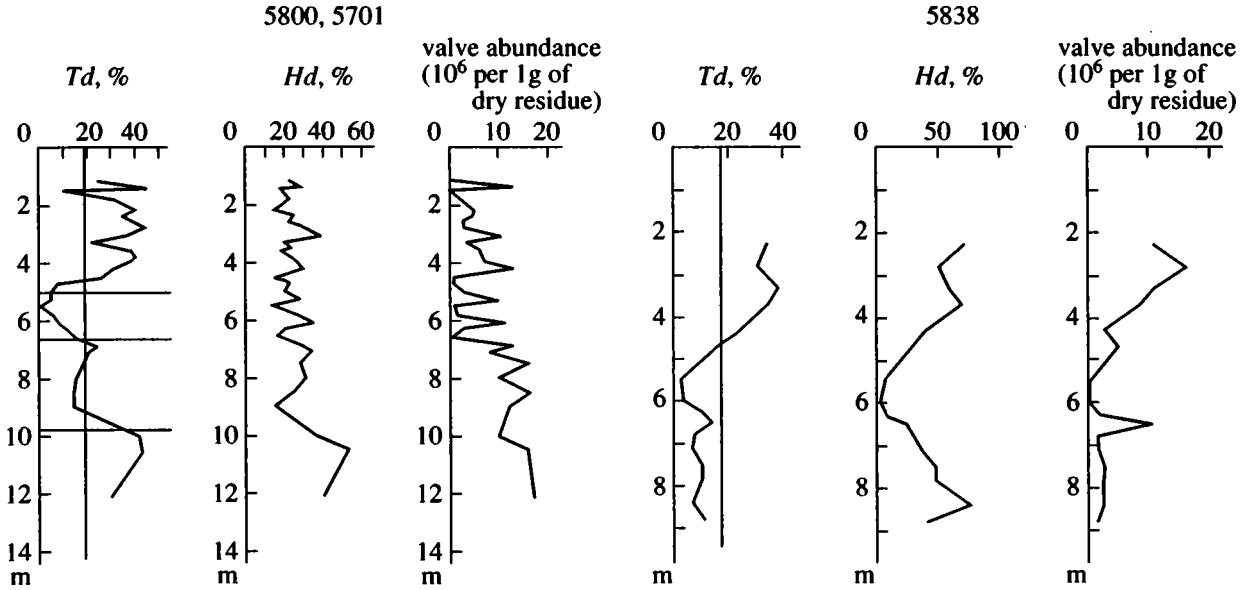


**Fig. 6.** Ecological types and climatic series of diatoms (W1 and W2—warm-water series and their ordinal number; C-W—moderately warm-water series; C—cold-water series).

*asteromphalus, C. radiatus, C. perforatus, Actinocyclus divisus, Thalassionema nitzschioides, Cyclotella striata, Hyalodiscus obsoletus, Navicula marina*, and *Lyrella lyra* (Jouse, 1962; Koizumi, 1973).

The *Xc* group is represented by *Thalassiosira gravida* var. *gravida, Th. hyalina, Th. kryophila, Th. antarctica, Th. nordenskioeldii, Bacterosira fragilis, Coscinodiscus marginatus, C. oculus–iridis, Rhizosolenia hebetata* f. *hiemalis, Thalassiothrix longissima*, and *Neodenticula seminae*.

Using the Td indices, we are able to distinguish the warm- and cold-water assemblages (Figs. 3–6), which can be combined into various climatic series. As a rule, the thermal and facial characteristics of assemblages change sumultaneously. In order to characterize the relative depth of dwelling and facial affinity of diatom assemblages, we suggest using the analogy with Td to calculate the depth coefficient:

$$Hd = 100Xp/(Xp + Xb),$$

where *Xp* and *Xb* represent the valve abundance of planktonic and benthic diatoms, respectively.

The climatic series of diatoms from sites 5701 and 5800 consists of four successive assemblages (Figs. 3 and 6) reflecting paleoclimatic changes and transgressive–regressive cycles of sedimentation.

*Assemblage I* (Td =43%) demonstrates that the climate of that time was warmer than at the present time. In the region under consideration, the present-day Td value is 16–17%, and the annual isotherm is 7°C. The mean annual temperature of surface water in the past was up to 5–6°C higher than now. According to Hd = 35–52%, the depth range of the assemblage corresponded to that of the intermediate shelf zone.

*Assemblage II* (*Td* up to 14–15%) is indicative of climatic cooling at that time. The depth range of this assemblage (*Hd* to 25–35%) was 40–50 m.

*Assemblage III* (*Td* = 1.5–10%) evolved under climatic environments colder than at this time. As *Hd* values indicate, the depth range of the assemblage was 40–50 m like before.

*Assemblage IV* (*Td* = 50%) is the most thermophilic as compared to others, but its *Hd* parameters remain unchanged. Judging from *Td* values characteristic of subtropical latitudes (Koizumi, 1986), the mean annual temperature of surface waters in this period could be up to 8–10°C higher as compared to the current state. The high Td values of the last assemblage and low magnitudes of this parameter in assemblages *II* and *III* seem to be controlled by the Kuroshio Current migration. At the time of assemblages *I* and *IV*, the current migrated to the northwest, closer to the South Kuril Islands.

The section in question reveals a transgressive sedimentary series. The transgression was at the maximum during the second peak of warming (assemblage *IV*) after a short-term stage of stable or slightly decreased sea level (assemblages *II* and *III*).

Similar diatom assemblages were also distinguished at Site 5838 (Figs. 4 and 6). The analogy of diatom climatic series at the sites 5800 (5701) and 5838 is evident from the paleoecological succession of fossil assemblages and from their index species. The diatom assemblages *I–III* from Site 5838 recorded the same paleogeographic events as assemblages *II–IV* from Site 5800 (5701) did. Assemblages *I* and *II* of a moderately cold-water type correspond to the short-term phase of the transgression delay, whereas the most thermophilic assemblage *III* reflects the second, maximum peak of

this event. A high proportion of neritic and oceanic species (up to 75%) suggests the diatom origin from the depth below the isobath of 50 m.

The diatom assemblage from Site 5837 (Figs. 5 and 6), displaying a high abundance of *Paralia sulcata* var. *biseriata*, appears to correspond to a later transgressive stage. Judging from ecological niches of *Paralia sulcata* et vars. (Jouse, 1962), this assemblage originated from the depth of about 20 m. The shallow-water sediments of Site 5837 grade southward, along the bed strike, into the deep-water facies exposed at Site 6095 (Fig. 2), where a high abundance of planktonic (to 65%) diatoms in assemblage *II* suggests their origin in a zone located beyond the isobath of 50 m.

Data on silicoflagellate distribution support the inference based on diatom analysis. In all sections, except that of Site 5837, we identified *Paradictyocha polyactus* Freng. f. *completa, Distephanus speculum* var. *speculum, D. speculum* var. *septenarius, D. octonarius* var. *polyactus*, and *Dictyocha fibula* var. *fibula*. All these species, except *D. speculum*, are of the warm-water type (Jouse, 1969). The limiting temperature is 16°C for *Paradictyocha polyactus* f. *completa* and 18–20°C for *Dictyocha fibula* var. *fibula* (Jouse, 1969). The usual temperature of *Distephanus speculum* evolution is about 0°C, and its association with the above species in the studied sediments is a consequence of their seasonal development and dominance. At the same time, it is clear that sediments of all sections are of the interglacial type, and predominance of *Dictyocha fibula* var. *fibula* and *Paradictyocha polyactus* f. *completa* (2–3% in sum) in assemblages *I, IV* (sites 5800, 5701), *III* (Site 5838), and *II* (Site 6095) indicates that the respective climatic phases were warmer than now. On the other hand, these species are scarce in assemblages *II, III* (sites 5800, 5701), *I, II* (Site 5838), and *I* (Site 6095), where the abundance of *Distephanus* forms grows up to 1–2%. Consequently, these assemblages mark the short-term cooling phase that took place during the interglacial period.

Spores and pollen are extremely rare in the studied deposits. The spore–pollen assemblage from the middle part of the section at site 5800 (5701) shows a predominance of arboreal pollen and consists of the following taxa: *Picea* sect. *Eupicea* (33.7%), *Pinus* cf. *koraiensis* (17.6%), *Quercus* (9.2%), *Pinus* s/g *Diploxylon* (5.2%), *Abies* (5.9%), *Betula* (4.8%), *Viburnum* (0.7%), *Alnus* (0.7%), *Ulmus* (1.6%), *Corylus* (0.6%), *Philodendron* (0.7%), *Juglans* (0.7%), *Aralia* (1.0%), *Tilia* (0.3%), *Carpinus* (0.9%), and *Fraxinus* (1.3%). Gramineae (29.7%), Compositae (22.5%), and *Artemisia* (21.6%) are dominant among the herbaceous taxa, and Polypodiaceae (91.2%) preponderate amid spores. According to conclusion of N.B. Verkhovskaya (Institute of Biology and Soil, Far East Division of Russian Academy of Sciences), this spectrum reflects the climatic conditions comparable with the current environments.

The spore–pollen assemblage from Site 5838 includes *Picea* sect. *Omorica* (up to 39.6%), *Betula* sect. *Costatae* (up to 17%), *B.* sect *Albae* (up to 29%), *Ulmus* (up to 14%) *Quercus* (up to 13%), *Juglans* (up to 10%), *Tilia* (up to 5%), *Carpinus* (up to 3%), *Fraxinus* (up to 6%), and *Fagus* (up to 1%). Gramineae (up to 80%) are dominant in the herbaceous group, and Polypodiaceae (up to 90%) are most abundant amid spores. This spectrum, corresponding to mixed coniferous–broad-leaved forests, reflects the warm climatic environments of the middle Pleistocene interglacial. The spore–pollen assemblage of Site 5837 differs somewhat from that of Site 5838. The dominant pollen of its arboreal group is that of *Abies, Picea* sect *Eupicea, Betula* sp., and *Alnus*, whereas the broad-leaved specimens, among which there is pollen of *Fagus* and *Tilia*, are less abundant and diverse in comparison with these taxa from the last site. This spectrum, when compared with those from sites 5800 and 5838, apparently corresponds to a cooler climatic phase.

## AGE AND CORRELATION OF DIATOM ASSEMBLAGES

Identifying the age of the studied sediments, we used the known dates of the diatom levels in the North Pacific sediments (Koizumi and Tanimura, 1985; Koizumi, 1986; Barron, 1992; Barron and Gladenkov, 1995). The stratigraphic subdivisions of these sediments are correlated with units of the Plio–Pleistocene diatom zonation accepted for the North Pacific region (Koizumi and Tanimura, 1985) and further detailed in its Pleistocene interval (Pushkar and Cherepanova, 1995; Korotkii *et al.*, 1997). Rhythms of paleoclimatic events were compared with those revealed by data of the oxygen-isotope stratigraphy (Williams *et al.*, 1988).

Amid diatoms the age indicators are species *Simonseniella barboi, S. curvirostris, Pyxidicula dimorpha, Actinocyclus ochotensis* var. *fossilis, Thalassiosira nidulus* var. *nidulus, Th. jouseae*, and *Th. gravida* var. *fossilis* extinct during the middle and initial late Pleistocene. *Pyxidicula zabelinae, P. schenckii, Coscinodiscus pustulatus, Neodenticula koizumii*, and *Thalassiosia antiqua*, all characteristic of Pliocene assemblages from Kamchatka and Sakhalin (Oreshkina, 1980; Gladenkov, 1988, 1994), and the northwest Pacific (Kozumi, 1973), are almost missing from our assemblages; this suggests that the studied diatoms were buried *in situ*. Simultaneously, *Neodenticula seminae, Simonseniella curvirostris*, and *Pyxidicula dimorpha*, whose origin marks the Plio–Pleistocene boundary, associate in our assemblages with diverse *Thalassiosira* forms and point to the Pleistocene age of host sediments.

The lack of extinct Eopleistocene–earliest Pleistocene species *Actinocyclus oculatus, Nitzschia reinholdii, Neodenticula koizumii*, and *Coscinodiscus marginatus* var. *fossilis* is an argument to attribute the distinguished diatom assemblage to the *Simonseniella*

*barboi* Zone (0.55–0.28 Ma). According to the preliminary paleomagnetic data of V.V. Shapovalov (Pacific Instute of Oceanography, Far East Division of Russian Academy of Sciences), the upper members of the Golovnino Formation show the normal polarity corresponding to the Brunhes Chronozone. In the shelf sediments of eastern Sakhalin, the upper boundary age of this zone ranges from 272000 ± 14000 (K-244) to 280000 ± 13900 years (K-245), and its lower boundary is 550000 ± 23100 years old according to results of the thorium–uranium isotope dating (Korotkii *et al.*, 1997). The respective age interval corresponds to the full chronostratigraphic range of the *Simonseniella curvirostris* Zone (Koizumi and Tanimura, 1985).

The revealed analogy between the climatic series of diatoms from the studied localities and the Sakhalin shelf allows us to attribute them to the Mindel–Riss Interglacial, which occurred between 430 and 280 (313) thousands years ago. In addition, the climatic series of sites 5800 (5701) and 5838 correspond to the climatic rhythms of the oxygen isotope stage (OIS) 11. At sites 5837 and 6095, they are correlative with the rhythms of OIS 9.

The analyzed distribution of silicoflagellates led us to a similar conclusion. The found specimens of *Distephanus octonarius* var. *polyactus* indicate the Mindel–Riss age of host deposits, because Jouse (1969) demonstrated that the extinction level of this form is a valid marker of the Mindel–Riss Interglacial in the North Pacific. The last occurrence level of *Paradictyocha polyactus* f. *completa* also seems to be at the end of this interglacial. In any case, its distribution area was displaced at that time into the Gulf of California (Jouse, 1969). One more indicative form *Dictyocha fibula* var. *fibula* disappeared from the high-latitude ecologic communities after the same interglacial epoch as well (Jouse, 1969).

## MAGNITUDE AND CORRELATION OF THE LATE GOLOVNINO TRANSGRESSION

The Otobetsugawa, Hayakita, and Dateyama formations in Hokkaido (Kitagawa *et al.*, 1987) are possible equivalents of deposits left by the Golovnino transgression. In the Plio–Pleistocene sections of the Oga Peninsula (Honshu), the *Simonseniella barboi* Zone presumably corresponds to the upper part of the Wakimoto Formation coupled with the Shibikawa Formation, because both units yielded the diatom assemblage comparable with the zonal one (Koizumi, 1977).

The diatom assemblage of *Simonseniella barboi* Zone was recovered from the Yanrakinot Horizon of marine deposits in the eastern Chukchi Peninsula (Pushkar and Roof, 1995), where this unit is dated back to the Mindel–Riss Interglacial. In western Alaska, this zone was distinguished in the Cape Blossom Formation of the Baldwin Peninsula (Pushkar and Roof, 1995) corresponding to OIS 11 and marking the time of the Anvilian transgression (Huston *et al.*, 1990; Kaufman *et al.*, 1993). The Wainwrightian transgression in the Arctic coastal region and the diatom assemblage of *Simonseniella barboi* Zone were discovered here in the marine Karmuk Beds (Pushkar and Roof, 1995).

It is estimated that the coast of Western Alaska experienced the glacioisostatic uplift during the last 300–400 thousands years and rose up to +13 m. In this case, the true top level of the Anvilian deposits is at the altitude of +10–12 m, and the same bench mark relative to the present state is suggested for the sea level of that time (Kaufman *et al.*, 1993). Bearing in mind the dwelling depth of *Paralia sulcata* var. *sulcata* (about 20 m), we may assume that the sea level at the formation time of the Cape Blossom Formation, whose top is at the altitude of +15 m (Huston *et al.*, 1990), was 22–25 m higher than now.

If we take into account these bench marks and the dwelling depth greater than 50 m inferred for the diatom assemblages from sites 5838 and 5800, then we may calculate the rate of tectinic uplift of the Kunashir Island during the last 430 thousands years (middle Pleistocene) was 0.28 mm/year. This value is one order of magnitude less than that calculated by Melekestsev *et al.* (1974), but very close to the tectonic rate values (0.3–0.4 mm/year) estimated in other works for this island (Fedorchenko and Piskunov, 1974) and Hokkaido (Okada and Ikeda, 1991). Unfortunately, strict criteria for judging the rate of tectonic uplift of the Kunashir Island are unknown for the Pleistocene time, and it is impossible to identify the true position, undisturbed by tectonics, of the Golovnino Formation top that is necessary, in addition to paleoecological data, to calculate the magnitude of the Golovnino transgression. The results presented below are hypothetical and have been obtained by extrapolation of data from the northern Hokkaido (Okada and Ikeda, 1991) and Alaska (Kaufman *et al.*, 1991, 1993).

The formation period of deposits at Site 5800 (5701) spans two transgressive peaks of OIS 11. Judging from several parameters, such as the ecological type of diatom assemblages, the altitude of the Golovnino Formation roof, and the rate of tectonic uplift of the island, the maximum sea-level stand could be as high as +20–25 m. During the short-term regressive phase between the peaks, the sea level presumably was close to the current state. It is likely that the southern part of the Kunashir Island did not exist, and the strait between the island and Hokkaido was much wider. The Golovnin and Mendellev volcanoes represented the subsea volcanic structures (Markhinin, 1957). The regression of OIS 10 is marked in sections by the tuff accumulation and subsequent hiatus in sedimentation. During the accumulation time of deposits corresponding to OIS 9 (sites 5737 and 6095), the sea surface was close to the present level or +3–5 m above it. The shallow-water facies of deposits under consideration occur in marine terraces located at high hypsometric levels. For

instance, the transgressive phases of OIS 11 and 9 presumably were responsible for accumulation of beach facies exposed in terraces situated at the altitudes of 80–100 and 40 m, respectively. The inferred rhythms of sea-level fluctuations in the first half of the middle Pleistocene are correlative with the global peaks of transgressions and regressions (Machida, 1975; Gladenkov, 1988), but differ from them in magnitude.

Each of the discussed transgressive phases ended with a series of intense volcanic eruptions marked by the accumulation of the tephroid members concordantly overlying the members of transgressive sedimentary facies.

## CONCLUSION

In the studied sections, we distinguished two transgressive–regressive cycles of sedimentation of the late Golovnino time. The transgressive bed series are represented by shelf facies deposited at different depth levels. Each of the transgressive phases of sedimentation was marked at the end by events of significant volcanic activity responsible for accumulation of thick pumice-bearing tephroids and subaerial tuff beds, which rest on the marine sedimentary strata. The Golovnin Volcano and the volcanoes situated in Hokkaido were the source structures of volcanic material. The former mainly served as the source of tuffaceous material, whereas some ash beds of the rhyolite composition enriched in potassium were related in origin with the second group of volcanoes.

Data of the diatom analysis were used to refine the age of the top layers of the Golovnino Formation. The upper members of this formation accumulated at the time of OIS 11–9 corresponding to the upper portion of *Simonseniella barboi* Zone of the diatom zonation. This conclusion is supported by data on distribution of silicoflagellates and by correlation of the deposits in question with the diatom levels well dated in the East Sakhalin shelf, Chukchi Peninsula, and Alaska. As a paleogeographic event, the late Golovnino trensgression is correlative with the Anvilian transgression in Alaska and the Yanrakinot transgression in the Chukchi Peninsula.

The comparative analysis of altitudes marking the upper boundary of the middle Pleistocene deposits and the estimated rate of tectonic uplift in the region allow us to conclude that the sea level was +20–25 m higher than now during OIS 11, and about 5 m higer at the time of OIS 9. During the regressive phase of OIS 10, the sea shelf was at least 110 m deep.

Apparently, the Golovnino Formation shoud be considered as a series of beds spanning the stratigraphic interval from the upper Pliocene to the and of the middle Pleistocene, inclusive, and accumulated in epochs contrast in the paleogeographic and tectono-volcanologic aspects.

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Reviewer Yu. B. Gladenkov

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# The History of Circulation in the Sea of Okhotsk during the Late Pleistocene–Holocene as Evidenced by Foraminifers

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**Abstract**—Changes in composition of planktonic foraminifers in nine cores of bottom sediments obtained from different areas of the Sea of Okhotsk were analyzed in order to reconstruct the history of climate and circulation of the sea, and the influence of adjacent basins on the sea hydrology. The revealed changes in composition of benthic and planktonic foraminiferal assemblages throughout the cores showed that the Sea of Okhotsk hydrology recurrently changed during the last 60 ka. During this period, the temperature of surface and bottom waters in the central part of the Sea of Okhotsk was generally lower than at present and close to that characteristic now of the northern part of the sea. The warming episodes of surface and bottom waters occurred 50, 30, and 13.280 ka ago. Peculiarities in distribution of thermophilic planktonic species *Globorotalia scitula* appear to be related to changes in the regime of the Soya and Tsushima currents. Lower parts of the cores barren of this species and corresponding to the time interval of 60 to 13 ka suggest the lack of the warm water inflow from the Japan Sea. The appearance of *G. scitula* in cores at the level of 13 ka indicates the renewal of the activity of the Soya and Tsushima currents. Changes in the regime of these currents are related to sea-level fluctuations.

**Key words:** *foraminifers, paleocirculation, upper Pleistocene, Holocene, Sea of Okhotsk*

Sea-level changes that substantially influenced the general circulation in oceans and the water exchange with marginal seas represent important phenomena in the Late Quaternary climatic history. Marginal seas are characterized by high sedimentation rates, and very brief climate-related events are recorded in their sedimentary sequences.

We analyzed changes in the composition of planktonic foraminifers in nine cores of bottom sediments obtained from different areas of the Sea of Okhotsk in order to reconstruct the history of climate and circulation of the sea and the influence of adjacent basins on the sea hydrology.

Changes in the composition of different faunal and floral groups such as diatoms, pollen and spores, benthic and planktonic foraminifers, and radiolarians against the background of changes in lithology and $CaCO_3$, $SiO_2$, and $C_{org}$ contents in Quaternary sediments have been considered earlier (Bezrukov and Lisitsyn, 1957; Koreneva, 1957; Saidova, 1961; Jouse, 1962; Kurihara, 1982). These investigations revealed synchronous changes in all the listed parameters in different areas of the sea and allowed these changes to be related to the glacial–interglacial environments. Nonetheless, the sporadic sampling and absence of absolute dates confused the relations between these changes and certain climatic events. None of these studies concerned the problem of the water exchange between the Sea of Okhotsk and both the Pacific and the Sea of Japan.

## MATERIALS AND METHODS

Foraminifers were studied in nine cores obtained from the Akademii Nauk and Instituta Okeanologii undersea rises, Deryugin Basin, and Kamchatka Peninsula slope, during Leg 53 of the R/V "Vityaz" (Fig. 1, Table 1).

Benthic and planktonic foraminifers were studied under the microscope MBI-2 in granulometric fractions of >0.25, 0.25–0.1, and 0.1–0.05 mm. The total abundance of shells of both foraminiferal groups was calculated per 1 g of sediment. Proportions of both benthic and planktonic species were estimated to characterize the structure of assemblages.

## MODERN HYDROLOGY OF THE SEA OF OKHOTSK

The Sea of Okhotsk is connected with the Pacific through a series of straits; some of them are more than 2000 m deep. Water masses of the Sea of Okhotsk are mostly formed at the expense of the Pacific waters entering the basin through the northern Kuril straits. The surface circulation is dominated by the cyclone current system covering almost the whole sea area. Several near-shore currents are most prominent: the warm Kamchatka Current, currents along northern and northwestern coasts, the stable East Sakhalin cold current passing in the north–south direction, and the relatively warm current from the Sea of Japan entering the Sea of Okhotsk via the Soya (Laperouse) Strait. The

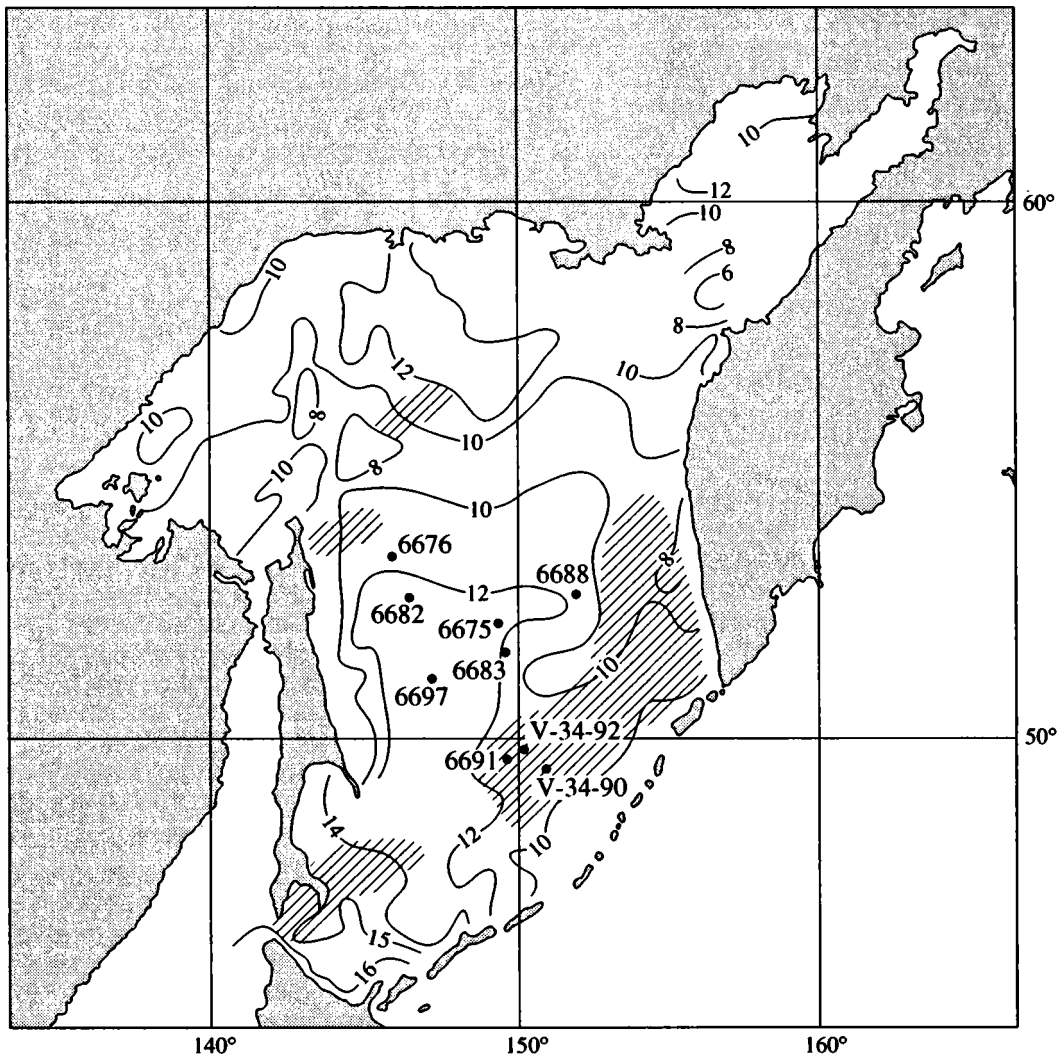


**Fig. 1.** Location of core sites and temperature (°C) of surface waters in the Sea of Okhotsk (generalized scheme after Moroshkin, 1966). Hatched are areas of the thermophilic phytoplankton distribution (after Kiselev, 1947).

Soya Current represents a branch of the warm Tsushima Current (Fig. 2). Its velocity amounts to 50–90 cm/s. According to calculations by A.K. Leonov (Ushakov, 1953), the total volume of waters transported to the Sea of Okhotsk by the Soya Current is 15000 km$^3$ per year. Such a significant inflow of waters from the Sea of Japan noticeably affects the temperature regime in the southwestern part of the sea and is responsible for the highest temperatures of surface waters near Hokkaido through all the seasons (averaging 12–14°C). The lowest temperatures (6–8°C, Fig. 1) are characteristic of areas along the Kuril Islands and of the northeastern part of the sea (Moroshkin, 1966).

The deep Pacific waters enter the Sea of Okhotsk through a gap in the ridge of the Kuril Islands (the Krusenshtern Strait area) at a depth of 800–1000 m to be spread in the sea within the depth interval of 600–800 to 1500 m. In the vertical water structure, they are distinguished as the Pacific deep-water mass characterized by high temperature (2.2–2.5°C), minimal oxygen content (0.5–0.8 ml/l), and high salinity (34.4–34.5‰). In the vicinity of the Instituta Okeanologii Rise, these waters raise up and occupy the surface layer. The Pacific deep-water mass is overlain by the intermediate water mass. The latter is formed by mixing of the Sea of Okhotsk and Pacific surface waters sinking in the Kuril straits and spreading over the entire Sea of Okhotsk within the depth interval of 300–700 m. These waters are characterized by temperatures of 1.5–2.0°C, which is substantially lower than at the similar depth in the western Pacific. The oxygen content in these waters is high (2.5–6.5 ml/l), and this differs them from both the intermediate waters of the ocean (0.5–5 ml/l) and the deep waters of the Sea of Okhotsk. Their salinity is up 33.3–33.9‰ (Moroshkin, 1966).

Characterstics of the studied cores

| Station | Coordinates | | Depth, m | Core length, cm | Location |
|---|---|---|---|---|---|
| | northern latitude | eastern longitude | | | |
| 6675 | 52°10.5′ | 149°20.3′ | 960 | 110 | Instituta Okeanologii Rise |
| 6675-2 | 52°10.5′ | 149°20.3′ | 960 | 288 | " |
| 6683 | 5149.5′ | 149°34.9′ | 1020 | 107 | " |
| 6691 | 49°29.0′ | 149°34.1′ | 976 | 610 | Akademii Nauk SSSR Rise |
| V-34-92 | 49°31.6′ | 150°10.2′ | 985 | 310 | " |
| V-34-90 | 48°49.9′ | 150°27.6′ | 1590 | 333 | " |
| 6676 | 53°31.4′ | 145°56.0′ | 1750 | 298 | Deryugin Basin |
| 6682 | 52°58.8′ | 146°08.2′ | 1766 | 257 | " |
| 6697 | 51°08.1′ | 147°11.9′ | 1360 | 735 | The central part of the sea |
| 6688 | 52°30.1′ | 152°00.0′ | 800 | 696 | Undersea slope of Kamchatka |

## LITHOLOGY AND AGE OF DEPOSITS

The studied cores are composed of silty mud with inclusion of pebbles, sand, ash material, foraminiferal and radiolarian shells, and diatom valves.

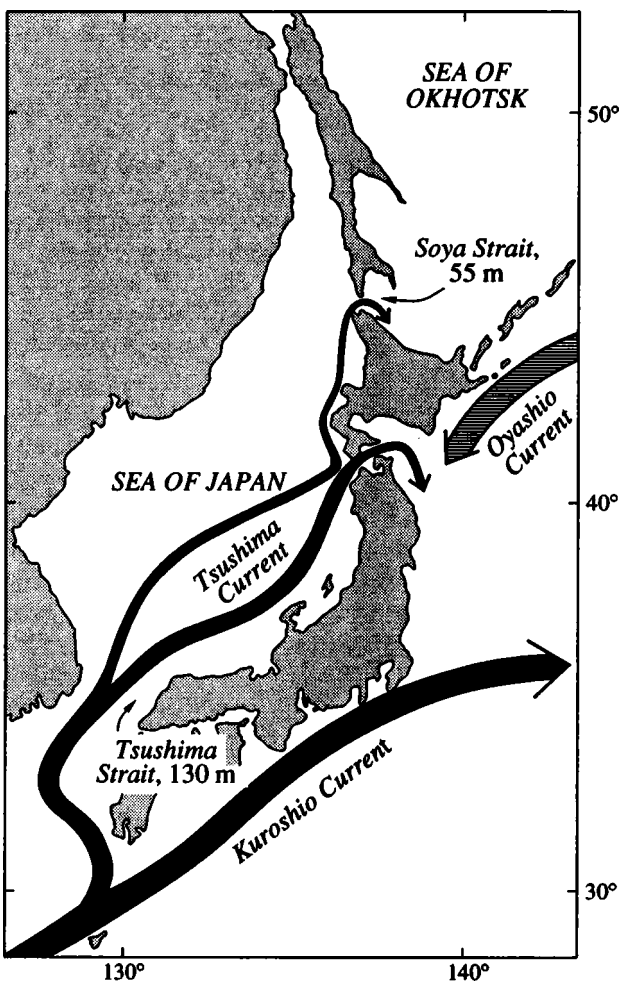


**Fig. 2.** Schematic surface-water circulation in the Sea of Japan (Oba *et al.*, 1991).

Most of the cores incorporate two sedimentary layers displaying a significantly higher content of foraminifers, radiolarians, diatoms, and other organisms than in other parts of the sections (Figs. 3, 4). The upper layer (I) was deposited during the last 12–13 ka (Gorbarenko *et al.*, 1988; Gorbarenko, 1991). The age of sediments from the base of the upper layer (Core V-34-90 from the Akademii Nauk Rise) is determined to be 13280 ± 80 years (Chekhovskaya *et al.*, 1995; Gorbarenko *et al.*, 1998).[1] According to Gorbarenko, the second layer (II) accumulated during the warm phase of the interstadial 3.3 dated back to 50–55 ka ago.

Sediments reflecting events of 24 and 30 ka ago were distinguished on the basis of climatic interpretation of our results (Belyaeva and Burmistrova, 1996, 1997), their comparison with the oxygen-isotope variation curve (Martinson, *et al.*, 1987), and the age of sediments from the base of cores estimated by using the average sedimentation rates.

Data on the modern distribution of benthic and planktonic foraminifers in the Sea of Okhotsk and World Ocean were used for paleooceanologic interpretation of changes in their assemblages through the cores (Parr, 1950; Phleger *et al.*, 1953; Bradshaw, 1959; Saidova, 1961; Todd, 1965; Lipps and Warme, 1966; Ingle *et al.*, 1980; Miller and Lohman, 1982; Arikawa, 1983; Peterson, 1983, 1984; Murray, 1984; Schnitker, 1984; Mead, 1985; Corliss *et al.*, 1986; Mackensen and Hald, 1988; Mackensen *et al.*, 1990; Loubere, 1991).

## DISCUSSION

The results of the study of the foraminifer assemblages in cores (Figs. 3 and 4) makes it possible to draw the following conclusion about changes in hydrology and deposition patterns in the Sea of Okhotsk during the last 60 ka.

[1] The radiocarbon dating was performed using the calcium carbonate shells of *Neogloboquadrina pachyderma* sin. by J.R. Southon at the Lawrence Livermore National Laboratory, California, USA.

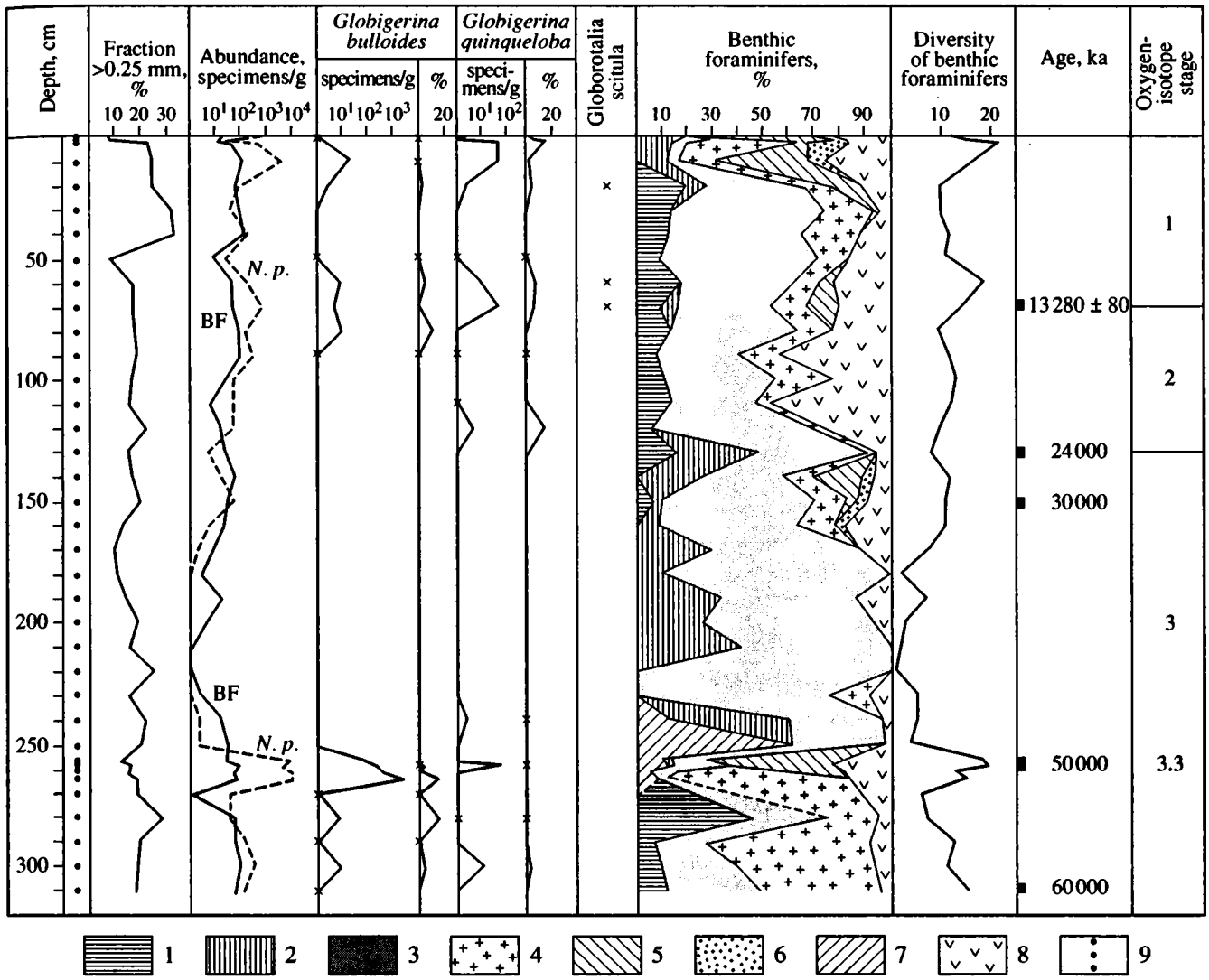


**Fig. 3.** Distribution of foraminifers in core V-34-92.
(*N. p.*) *Neogloboquadrina pachyderma*; (BF) benthic foraminifers; (×) single species shells in the sample. Benthic species: (1) *Angulogerina angulosa*; (2) *Uvigerina peregrina*; (3) *Uvigerina auberiana*; (4) *Cassidulina teretis*; (5) *Cassidulina laevigata*; (6) *Valvulineria ochotica*; (7) *Martinottiella communis*; (8) other species; (9) studied samples.

During the period from 60 to 50 ka, the surface water conditions in the central part of the sea differed from those observed there nowadays. The planktonic foraminifer assemblage was dominated by *Neogloboquadrina pachyderma* sin. (Ehrenberg) (84–100%) accompanied by *Globigerina bulloides* d'Orbigny and *G. quinqueloba* Natland.[2] Contrary to the modern assemblage of the "central biofacies" of the Sea of Okhotsk, species *Globorotalia scitula* (Brady), *Globigerinita glutinata* (Egger), and *G. uvula* Ehrenberg are missing in the considered portion of the sequence. Conditions near the bottom also differed from those characteristic at present. The absence of very small thin-walled and less dissolution-resistant shells of *G. uvula* and *G. glutinata*, characteristic of the "central biofacies," and shells of small *G. bulloides* in combination with the low content of small *N. pachyderma* suggest a lesser saturation of bottom waters with calcium carbonate in the past than at present. The preponderance of *Cassidulina teretis* Tappan in association with *Angulogerina angulosa* Williamson in the benthic foraminifer assemblage indicates the low temperature, high hydrodynamics, and organic matter influx to the bottom sediments. The elevated content of *C. teretis* during the cooling periods was noted in the upper Pleistocene sed-

[2] The ratio of sinistral to dextral forms of *N. pachyderma* is not considered, because the sinistral forms compose more than 98% and, usually, 100% of the assemblage from the surface sediments (Lipps and Warme, 1966) and, according to our and Kurihara's (1982) estimates, from the core samples recovered in the Sea of Okhotsk.

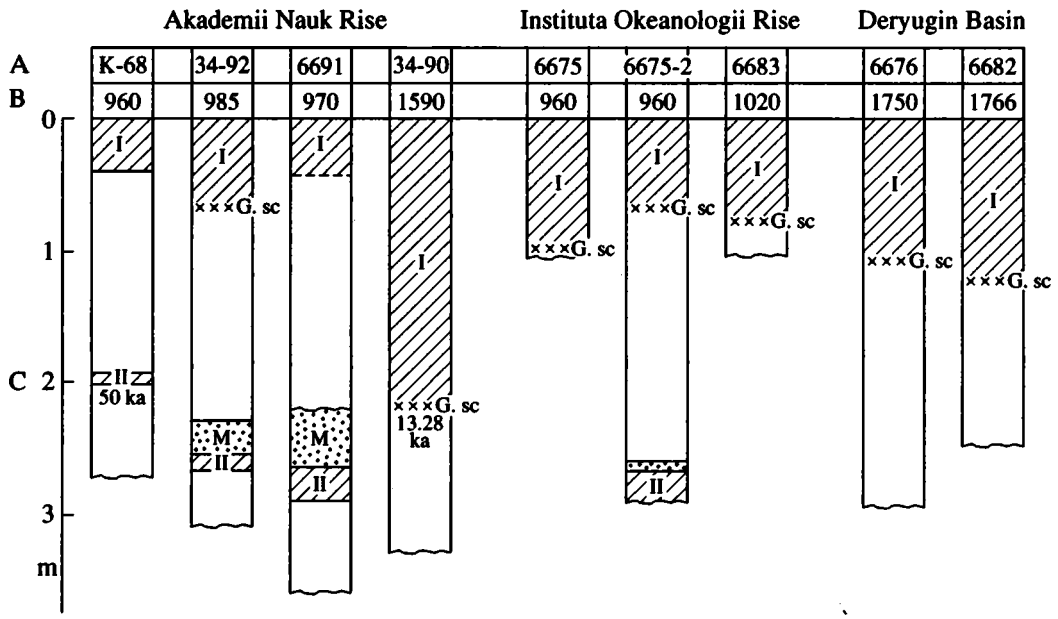


**Fig. 4.** Distribution of *Globorotalia scitula* and *Martinottiella communis* in cores from the Sea of Okhotsk.
(A) site number; (B) depth; (C) core length; (I, II) sediment layers enriched in organic remains: (G. sc.) first occurrence level of *Globorotalia scitula* in cores; (M) sediments with the high content of *Martinottiella communis*.

iments of many areas of the North Atlantic (Feyling-Hansen *et al.*, 1983; Murray, 1984; Schnitker, 1984).

Approximately 50 ka ago, the hydrologic situation changed drastically. Sediments corresponding to this time are characterized by a sharp increase (by one–two orders) in the planktonic foraminifer abundance, and by the presence of large (more than 0.25 mm) well-developed and abundant small shells of *N. pachyderma* and *G. bulloides*, which indicate the elevated temperature of surface waters and the growth of planktonic foraminifer productivity. The sharp decrease in the share of cold-water species *C. teretis* and the prevalence of *Cassidulina laevigata* d'Orbigny peculiar now of the Pacific deep waters among the benthic foraminifers can be explained by the significant warming of bottom waters. The assemblage with preponderant *C. laevigata* includes 19 species, all of whose specimens are well-preserved. Sediments of this time period are correlative with the warm climatic event corresponding to the oxygen-isotope phase 3.3 (Martinson *et al.*, 1987; Gorbarenko *et al.*, 1988).

It is remarkable that, in all the cores recovered from the rises, this layer enriched in organic remains is overlain by the layer with abundant agglutinated species *Martinottiella communis* (d'Orbigny), later which became extinct. The horizon of the *M. communis* prevalence can serve as a reliable marker for tracing the 50-ka-old level in the cores from rises of the Sea of Okhotsk.

Upsection, in the interval corresponding to the time period of 50 to 30 ka ago, the planktonic foraminifer assemblage is represented only by rare or single large shells of *Neogloboquadrina pachyderma*, one of the most cold- and dissolution-resistant species. This "assemblage" is similar to the modern assemblage of the "coastal biofacies" in the northern part of the sea, where single specimens of *N. pachyderma* occur only at separate sites. It is apparent that, in the period from 50 to 30 ka ago, the surface water conditions in the central part of the basin were more severe, with temperatures close to those characteristic now of northern areas of the Sea of Okhotsk. The main factor controlling the fossil assemblage composition was shell dissolution. The benthic foraminifer assemblage shows an intense dissolution of shells in the course of their fossilization. The assemblage diversity equals four to five species in all samples. The most abundant are *Uvgerina peregrina* Cushman and *U. auberiana* d'Orbigny associated with subordinate *Cassidulina teretis*. The first two species are characteristic representatives of the foraminiferal fauna from highly productive areas of the World Ocean (Miller and Lohman, 1982; Corliss *et al.*, 1986; Loubere, 1991).

Only large (more than 0.25 mm) shells are preserved in benthic foraminifer assemblages influenced by strong dissolution. The persistent presence of *U. peregrina* and *P. auberiana*, as well as their relatively high abundance in some samples against the background of strong dissolution and dilution by the terrigenous material, allow us to suggest that, during the accumulation of these sediments, the surface waters were characterized by a high bioproductivity which resulted in a high organic matter influx to the sea floor. The bottom water circulation was probably weak, and the decomposition of a great mass of organic matter

resulted in an elevated carbonic acid content in the interstitial water of the bottom sediments. Taking into account the strong dissolution effect, the temperature regime near the bottom can hardly be interpreted. Nonetheless, the faunal changes at the base of deposits with the dissolved foraminifer assemblage suggest that the bottom waters were characterized by low, probably negative temperatures.

Using the oxygen-isotope curve by D. Martinson, we date the noticeable warming that occurred later at 30 ka. It is recorded in all the cores by the increase in the planktonic foraminifer abundance and by the appearance of *Globigerina bulloides* and *G. quinqueloba*, whose content among foraminifers in this part of the core is similar to that in the modern biocoenosis of the studied area (Lipps and Warme, 1966). The share of young forms also increases here. The more active circulation and higher temperatures of bottom waters are evident from the appearance of *Angulogerina angulosa* and *Cassidulina laevigata* among the benthic foraminifers. The respective period is also marked by a better preservation of shells of both planktonic and benthic foraminifers.

This warming was replaced by a sharp cooling 24 ka ago at the boundary of the 2nd and 3rd oxygen-isotope stages. In the cores studied, this event is reflected by the disappearance of thermophilic *C. laevigata* and by prevalence of cold-resistant species of benthic and planktonic foraminifers, as well as by a better preservation and by the diversity increase of benthic foraminifers. The bottom water conditions were similar to those in the northern part of the sea at the depth exceeding 200 m and to those assumed for the formation time of deposits from the basal part of the cores in question. These cold conditions existed till the time of 13280 years ago.

The most drastic change in the faunal assemblages and environments occurred 13280 years ago, i. e., at the boundary between the 1st and 2nd oxygen-isotope stages.

In general, the period of formation of upper sediments recovered by cores from the central part of the Sea of Okhotsk was characterized by a decrease in the terrigenous dilution and dissolution intensity. The planktonic foraminifer assemblage from the upper part of the cores is most diverse. For instance, cores from the Instituta Okeanologii Rise, Deryugin Basin, and slope of the Akademii Nauk Rise host all the species characteristic of the "central biofacies": *Neogloboquadrina pachyderma* sin., *Globigerina bulloides, G. quinqueloba, Globigerinita uvula, G. glutinata*, and *Globorotalia scitula*. The only core from the top of the Akademii Nauk Rise shows that dissolution occurred during the last 13 ka in this area against the background of low sedimentation rates. This is evident from the presence of rare fragments of shells and from the absence of *G. uvula* and *G. glutinata* species. Deposition of sediments during the last 13 ka proceeded under conditions close to those in the surface and bottom waters of the sea at present.

Sharp changes in the planktonic foraminifer abundance are observed against the background of stable composition of their assemblages. The highest abundance values are confined to the periods of maximum warming that occurred 11–13 and 10 ka ago. The minimal values probably correspond to the Younger Dryas (10–11 ka ago), when the bioproductivity of surface waters decreased. The decrease in bioproductivity is implied by the absence of the benthic *Uvigerina peregrina* species.

The growth in share of thermophilic deep-water species *Cassidulina laevigata* reflects the subsequent gradual elevation of the bottom temperatures. Changes in foraminiferal assemblages from the upper part of all the cores indicate a significant frequency of climatic oscillations that lasted about 1000 years and were sometimes of a substantial amplitude.

Only sediments from the upper part of the cores contain thermophilic planktonic species *Globorotalia scitula* (Figs. 3, 4). According to J. Southon (Gorbarenko *et al.*, 1998), the appearance level of this species in the core V-34-90 from the slope of the Akademii Nauk Rise is dated back to 13280 years.

The occurrence of *G. scitula* in modern sediments of the Sea of Okhotsk undoubtedly shows the relation to the warm Soya Current (Lipps and Warme, 1966). Before explaining the distribution peculiarities of this species in the cores, it seems reasonable to consider in greater detail the influence of the Soya Current on the formation of the modern fauna and flora of the Sea of Okhotsk. As mentioned above, the significant influx of warm waters from the Japan Sea affects the temperature regime of the southwestern part of the sea. The presence of thermophilic species in the phyto- and zooplankton communities, as well as in benthic assemblages of the southwestern and other areas of the Sea of Okhotsk, are related to the influx and further expansion of these warm waters (Shchedrina, 1953; Lubny-Gertsik, 1959; Ponomareva, 1959; and others). The distribution patterns of the water flow from the Sea of Japan (Fig. 1) are well illustrated by the distribution area of thermophilic phytoplankton forms (Kiselev, 1947). Thermophilic species of planktonic foraminifers occur in the Sea of Okhotsk as well (Shchedrina, 1953; Lipps and Warme, 1966). Subtropical species *Globigerinoides conglobatus* (Brady) and *G. ruber* (d'Orbigny) occur only in the southwestern Sea of Okhotsk. Species *G. scitula* characteristic of the transitional (after Bradshaw, 1959) Pacific fauna are more widespread in the recent sediments of the Sea of Okhotsk and detected on the Akademii Nauk and Instituta Okeanologii rises, in the Deryugin Basin, and on the Kamchatka slope. The northern boundaries of the distribution areas of *G. scitula* and thermophilic phytoplankton species coincide.

All the listed thermophilic planktonic foraminifers populate the Pacific surface waters only to the south of

40°N (Bradshaw, 1959). Such a position of the northern boundary of these thermophilic species preclude their migration into the Sea of Okhotsk via the Kuril straits. Thermophilic species from the Pacific may enter only the Sea of Japan using the Tsushima Current, which branches off the Kuroshio Current. In the Sea of Japan, the Tsushima Current is divided into several branches, one of which, forming the Soya Current, enters the Sea of Okhotsk. The occurrence of the Pacific-related thermophilic species in the Sea of Okhotsk can only be explained by their migration from the Sea of Japan via the Soya Current.

We relate the above-mentioned peculiarities in distribution of *G. scitula* in the studied cores to the past changes in the Soya and Tsushima currents. The absence of this species in the lower part of the cores corresponding to the time period of 60–13 ka allows us to suggest that there was no inflow of warm waters from the Sea of Japan at that time. This appears to be related to the sea level stand below the modern depth of the Soya Strait (55 m). The appearance of the species in sediments 13 ka ago indicates the sea-level rise and the resumption of the Soya Current. Based on the complete similarity in the distribution patterns of *G. scitula* nowadays and 13 ka ago, we suggest that principal characteristics of the Soya Current were close to the modern ones.

Our paleoecological interpretation is in accord with the results obtained for the Sea of Japan (Oba *et al.*, 1991). The study of fauna and flora from the Upper Quaternary sediments of the Sea of Japan accumulated during the last 85 ka revealed the absence of thermophilic foraminiferal, radiolarian, and diatom species in the lower part of the section corresponding, according to Japanese scientists, to the period from 85 to 11 ka. Oba *et al.* (1991), relate these distribution patterns of thermophilic species to the closure of the Tsushima Current at that time in response to the sea-level's fall down to the depth exceeding that of the Tsushima Strait (130 m). The appearance of thermophilic species of fauna and flora in the upper portion of the section is explained by the sea-level rise and resumption of water exchange between the Sea of Japan and Pacific via the Tsushima Strait.

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# Relationships and Age of the Lower Paleozoic Kursai, Karagaily, and Kirtabulga Formations in the Kastek Ridge (Zailiiskii Alatau)

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**Abstract**—The integral stratigraphic scheme for the Lower Paleozoic (Lower Cambrian to Middle Ordovician) is considered on the basis of new finds of organic remains in the Kursai (small shally fossils), Karagaily (algae, trilobites, brachiopods, conodonts, sponges, small shally fossils, and others), and Kirtabulga (trilobites, graptolites, conodonts, algae) formations. These finds furnished an opportunity to reliably correlate the individual sections, compile an integral regional stratigraphic scheme, substantiate the local biostratigraphic zonation, perform the interregional correlation, and confirm the autochtonous position of the Cambrian–Ordovician deposits in question.

**Key words:** *Northern Tien Shan, Zailiiskii Alatau, Cambrian, Ordovician, Kursai, Karagaily, and Kirtabulga formations, small shally fossils, algae, trilobites, conodonts, graptolites, Karagaily reef*

## INTRODUCTION

The Kursai, Karagaily, and Kirtabulga formations of the Early Paleozioc age are spread over the northern part of the Aktyuz-Boordy ore area in the southern periphery of the Muyunkum median massif. The main feature of the latter is the high-level position of the pre-Riphean basement, whose outcrops alternate with the superimposed Caledonian depressions and grabens. Over a significant area, the Caledonian and Precambrian structures are hidden under the quasiplatform cover of the Middle Paleozoic volcanogenic–terrigenous deposits.

D. I. Yakovlev was the first to mention the Lower Paleozoic rocks in this region. In 1937, this researcher sampled remains of Lower Cambrian algae and Upper Cambrian trilobites from the limestone sequence in the Karagaily River head and named this unit as the Karagailybulak Formation. In 1942, A.G. Vologdin sampled algae and archaeocyathid remains and also compiled the schematic columns of the lower Karagaily Formation. In 1962, V.I. Goncharova sampled the Upper Precambrian trilobites at the same sites.

The large-scale geological survey that commenced at the end of the 1950s allowed the Upper Paleozoic section to be studied only in general terms. For instance, N.I. Doroshenko and V.I. Kiselev divided the section into three units: the presumable Lower Cambrian Kursai Formation represented by basalts and terrigenous rocks; the Middle–Upper Cambrian Karagaily Formation consisting of sandy–shaly, siliceous–volcanogenic, and flyschoid subformations; and the Middle Ordovician Kirtabulga formation of flyschoid rocks incorporating large reef buildups. Owing to the block patterns typical of the area, the internal structure of these formations was established based on correlation of a number of particular sections. Until the recent time, their insufficient paleontological substantiation and lithologic affinity of separate fragments of heterochronous sequences prevented unambiguous correlation. This can be exemplified by the stratigraphic section of the Kiselev's Karagaily Formation beginning with the sandy–shaly subformation, in which we found organic remains of the Ordovician age.

Recently, the autochtonous position of the Karagaily Formation and Ordovician reef buildups, and, consequently, of the normal stratigraphic succession in the Lower Paleozoic section was questioned. The solution of these problems required that detailed biostratigraphic study be carried out in the region. The study was performed in two areas: in the Karagaily River head and its tributaries, and in the Kirtabulga River middle course. This resulted in new finds of algal remains, trilobites, brachiopods, conodonts, crinoids, graptolites, and small shally fossils (Fig. 1), which provided a possibility to correlate the isolated sections and to compile an integral Lower Paleozoic stratigraphic scheme for the time period from the Lower Cambrian to Middle Ordovician.

† Deceased.

## BIOSTRATIGRAPHIC RESULTS

The results of biostratigraphic studies of the Kursai, Karagaily, and Kirtabulga formations are presented below.

*The Kursai Formation* compose a tectonic block 2.5 km$^2$ in size that is located in the middle course of the Kirtabulga River. It is bounded by the Kirtabulga fault and by a system of submeridional dislocations. Separate outcrops of the formation were also observed in the Karagaily River head.

We revised and described the formation section in the stratotype locality in the Kirtabulga River basin. In addition to three members exposed there, we included into this unit the agglomerate basaltic tuffs and schistose tuff conglomerates referred by Kiselev to the Karagaily Formation in the Karagaily River basin.

At present, the formation is subdivided into four litholostratigrafic members representing two terrigenous–volcanogenic rhythms (Fig. 2).

*Member 1.* Lower layers of the unit are truncated by a fault, and its relationship with underlying deposits is unknown. The visible base of the member is represented by a horizon (about 100 m) of gray and brownish coarse-grained sandstone enclosing lenticular gritstone and small-pebbled conglomerate interlayers. These rocks are overlain with a sharp contact but without an apparent erosion by gray large-pebbled conglomerates (to 50 m). Pebbles in conglomerates are represented by various granitoids, pink and gray banded gneiss, migmatite, ophitic marble, and Precambrian actinolite schist. The member is crowned with a thick (to 250 m) horizon of gray medium-grained sandstone incorporating gray and lilac siltstone interbeds (5 to 50 m) with lenses of gray sandy limestone. The thickness of the member is 400 m.

*Member 2.* The unit overlies the sandstone member with a sharp even contact and is composed of dark green and black spheroid to pillow lavas of variolitic basalt. The member top is marked by a horizon (to 5 m) of gray–green banded chert. The thickness of the member is 200 m.

*Member 3* begins with a horizon (to 50 m) of gray small-pebbled conglomerates interbedding with gravely sandstone. Pebbles and rock clasts are represented by granite, gneiss, and rhyolite. The section is continued by poorly sorted sandstones (to 10 m), which are overlain by similar poorly sorted sandstone with gravely clasts of andesite (about 110 m). The sandstones incorporate gray siltstone interbeds with lenses of dark gray to black limestones. The chemical treatment of limestones revealed single remains of protoconodonts *Rhombocorniculum* sp. and sponge spicules of the *Protospongia* sp. type. The fossils allowed the host rocks to be attributed to the Atdabanian Stage of the Lower Cambrian. The member is crowned with a horizon of gray medium-grained sandstone with gritstone and small-pebbled conglomerate interbeds. The thickness of the horizon is about 70 m. The apparent thickness of the member amounts to about 240 m.

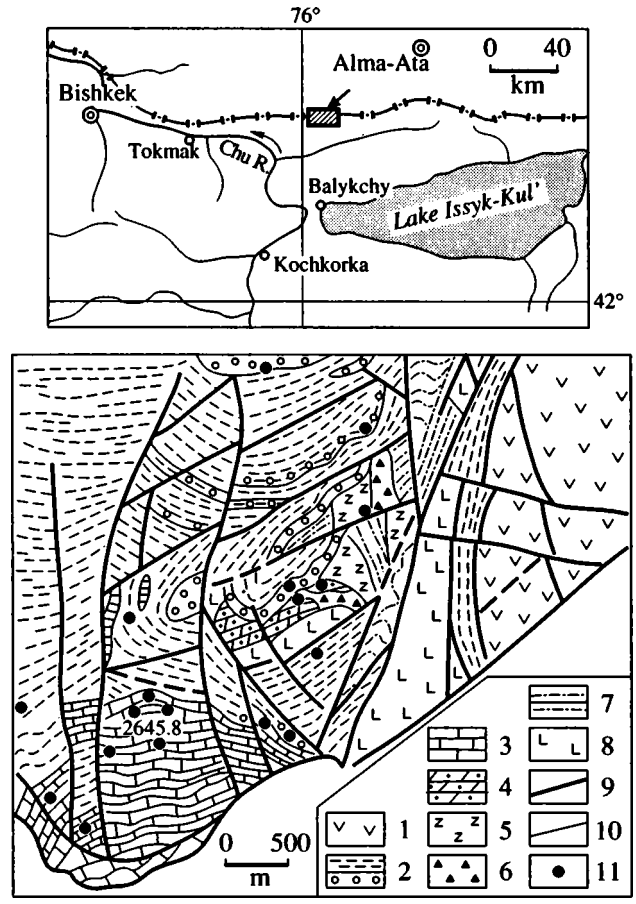



**Fig. 1.** The studied area and schematic geological map of Lower Paleozoic deposits of the Kastek Ridge: (1) volcanogenic deposits of the Middle Ordovician Oktorkoi Formation; (2, 3) the Lower–Middle Ordovician Kirtabulga Formation of (2) flyschoid deposits, (3) reefal limestones; (4–7) the Lower–Upper Cambrian Karagaily Formation divided into (4) *Member 1* (biohermal limestones), (5) *Member 2* (calcareous–siliceous), (6) *Member 3* (gray-colored flysch), (7) *Member 4* (variegated flysch); (8) the Lower Cambrian Kursai Formation (basalts and their tuffs, tuff conglomerates); (9) faults; (10) stratigraphic contacts; (11) sites of collecting of organic remains.

*Member 4.* The unit is exposed in the Karagaily River basin and composed of dark green basaltic agglomerate tuff and schistose tuff conglomerates with interbeds of tuffstone and tuffaceous siltstone. The top surface of the member is uneven, with pockets. The thickness is about 115 m.

The total visible thickness of the Kursai Formation is approximately 960 m.

*The Karagaily Formation* is exposed in a limited area of the Karagaily River middle course at the right-hand side of the Vodopadnyi, Kembriiskii, and Dzholdy creeks valleys. It overlies with the erosion surface the Kursai Formation. The unit is truncated by a sharp

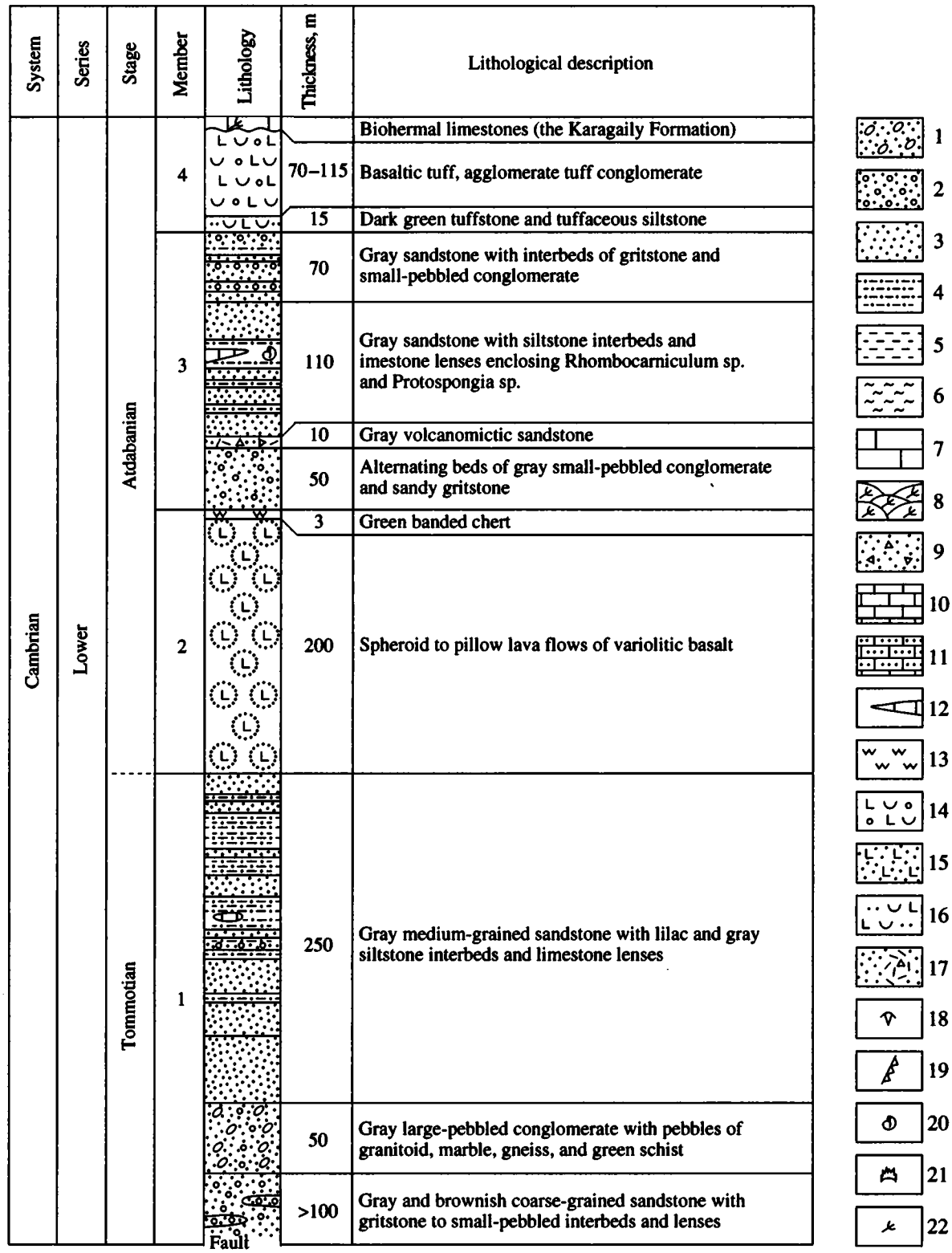



**Fig. 2.** The structure of the Kursai Formation at the left-hand side of the Kirtabulga River (the Kursai Creek) and in the Vodopadnyi and Kembriiskii creeks interfluve area. Lithology: (1) conglomerate, (2) gravel, (3) sandstone, (4) siltstone, (5) gray shales, (6) variegated shales, (7) thick-bedded limestone, (8) biohermal algal limestone, (9) interbiohermal conglobreccia, (10) platy limestone, (11) sandy limestone, (12) limy lenses and interbeds, (13) chert, (14) basaltic tuff and agglomerate tuffaceous conglomerate, (15) basaltic lava, (16) tuffstone and tuffaceous siltstone, (17) volcanomictic sandstone. Organic remains: (18) trilobites, (19) graptolites, (20) brachiopods, gastropods, and small shally fossils, (21) conodonts, (22) algae.

structural unconformity at the base of the conglomerates of the overlaing Kirtabulga Formation (Fig. 1).

As mentioned, our investigations showed that the lower sandy–shaly subformation (in Kiselev' s terms) does not belong to the Karagaily Formation. For instance, the greenish gray silicified shale, siltstone, and sandstone occurring in the upper part of this subdivision in the right-hand side of the Kembriiskii Creek yielded, as the chemical treatment revealed, rare remains of siliceous sponges of the Hexactinellida class and siliceous tubular problematics. Fossils similar to these problematics were recently described from the Lower Paleozoic deposits of inner Mongolia and from the northern areas of the North China platform (Liu Xiaoling *et al.*, 1990, 1992) as the Quinghezen and Changtu faunas. Fossils from our collection were determined by Mambetov as *Cylindrochites* cf. *daizhuanyziensis* Liu. Thus, the complex siliceous sponges (hexantinellids and pentactinellids) suggest that the age of the former "lower" subformation of Kiselev can not be older than the Ordovician. This conclusion is also supported by data on above-mentioned tubular problematics.

We divided the middle (sedimentary–volcanogenic) subformation (in Kiselev' s terms) into two parts. The lower, volcanogenic part was included, as previously mentioned, into the upper Kursai Formation (Fig. 2). It is separated from the sandy–shaly member by a distinct tectonic contact.

The top of the tuff conglomerate horizon is marked by isolated large pockets. Being filled with biohermal limestones, they look like limestone lenses inside conglomerates. Kiselev described them exactly in these terms. We established that some pockets are filled with breccia of the typical interbiohermal facies, but not with limestones. This breccia consists of limestone clasts closely spaced and embedded into the less abundant carbonate–tuffaceous cement. This type of filling excludes the possibility to consider the carbonate rocks as composing lenses, whereas the large pockets indicate an erosion episode before the deposition of biohermal limestones composing the base of the Karagaily Formation type section. Four member can be discriminated in the formation (Fig. 3).

*Member 1.* The unit is composed of biohermal limestones and interbiohermal breccia. Bioherms are characterized by distinct contours and sharp contacts; they have flat slopes subparallel to the bedding. Two complex chains of these buildups are separated by a thin horizon of redeposited tuffs with the carbonate cement. Some bioherms sized 50–200 × 25–35 m are separated along the chain strike by areas of similar dimensions occupied by interbiohermal breccia. Bioherms themselves are composed of gray to dark gray, massive, usually pelitomorphic and fine-grained, less commonly, white recrystallized (sugarlike) organogenic limestones. Locally, they show patchy coloration with dark gray or black irregular mottles 5–15 cm across. Interbiohermal breccia consists of angular clasts (5–15 cm) and large (0.5–0.7 m) blocks of biohermal limestones. The clastic material sharply prevails over the cement, which is carbonate–tuffaceous in composition.

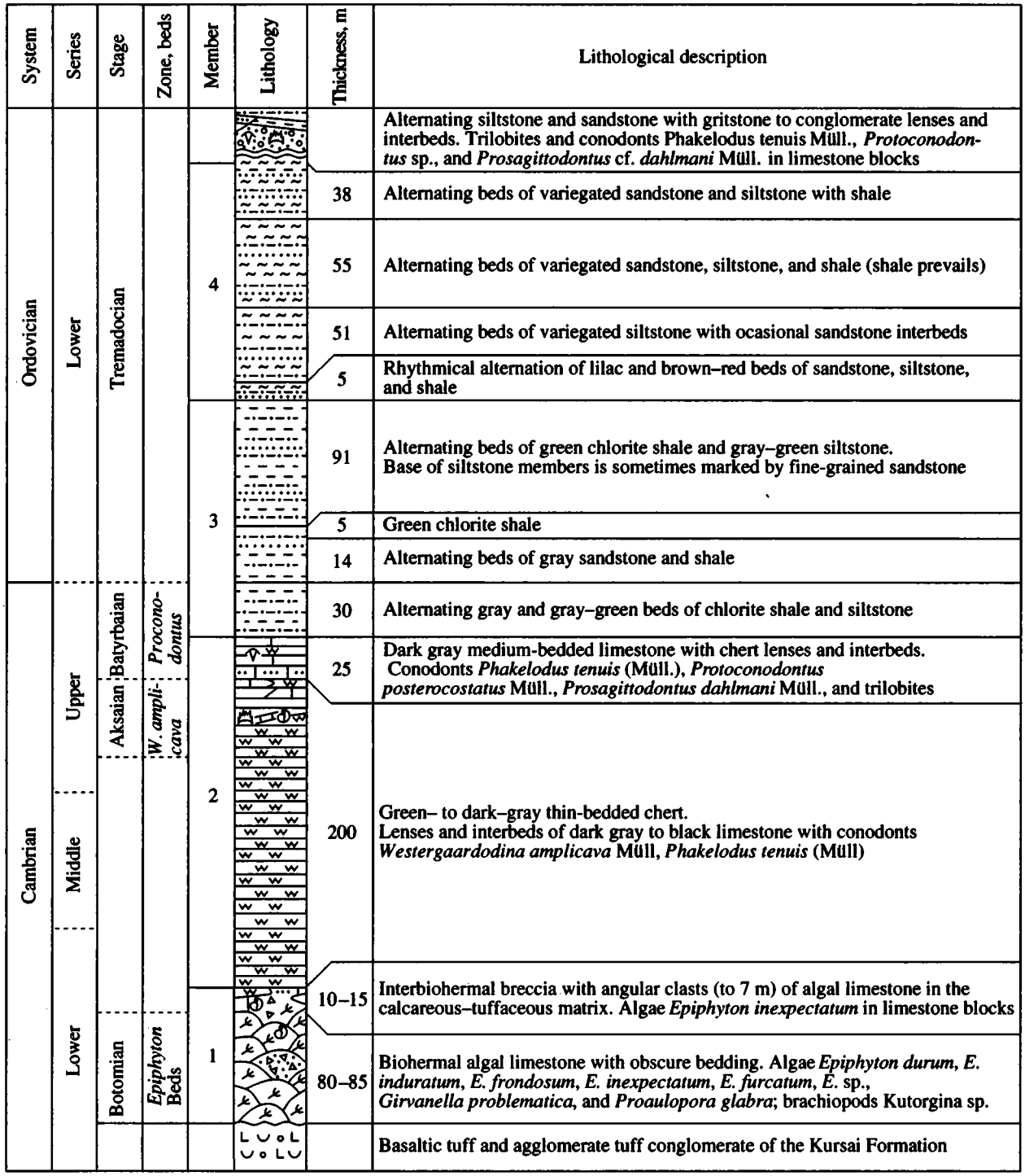
The basal layers of the lower biohermal massif yielded algae *Epiphyton durum* Korde, *E. induratum* Korde, *E. frondosum* Korde, *E. inexpectatum* Korde, *E. furcatum* Korde, *E.* sp., *Proaulopora glabra* Krasnop., and *Girvanella problematica* Nich. et Ether. According to V.A. Luchinina (Institute of Geology and Geophysics, Siberian Division, Russian Academy of Sciences), this algal assemblage is Early Cambrian in age.

From limestones of the upper biohermal body, Luchinina determined *Epiphyton inexpectatum* Korde and *Girvanella problematica* Nich. et Ether. The matrix of interbiohermal rocks includes the indeterminable trilobite remains, whereas the blocks yielded algae *Epiphyton inexpectatum* Korde.

The total thickness of the member is 80–85 m.

*Member 2* is composed of cherts (phtanites, phtanoids) and, to a lesser degree, of limestones. The transition between these rocks and underlying member is gradual. Upsection, the abundance and size of clasts in breccia of *Member 1* gradually decreases, and the rocks grade into carbonate–volcanomictic sandstone and siltstone. In upper horizons, first lenses and then interbeds of gray and green chert appear. The number of interbeds gradually increases, and terrigenous rocks become replaced by light gray, gray, and dark gray cherts sometimes displaying green shading. In the lower part, rocks are mostly thin-bedded, whereas upward, where they enclose interbeds of dark gray cherty shales and dark gray to black, sometimes sandy limestones, the thicker and obscure bedding prevails. Chert beds yielded the abundant sponge spicules of the Hexactinellida class and microproblematics. Limy layers enclose single indeterminable remains of trilobites, articulate and inarticulate brachiopods, and also conodonts determined as *Westergaardodina amplicava* Müll. and *Phakelodus amplicava* (Müll.). Based on conodonts, deposits are referred to the *Westergaardodina amplicava* Zone, which, in turn, allows the host rocks to be correlated with the topmost Aksaian Stage of the Upper Cambrian (Dubinina, 1982). The thickness of the substantially siliceous portion of the member is about 200–210 m. The member is terminated with a horizon (to 25 m) of dark gray medium-bedded, less commonly, thick-bedded limestones with interbeds of greenish gray thin-banded chert similar to that described above.

The chemical treatment of limestones revealed abundant remains of conodonts *Phakelodus tenuis* (Müll.), *Diaphanodus* sp., *Proconodontus posterocostatus* Müll.., and *Prosaggitodontus dahlmani* Müll. The assemblage dates this portion of the section about 25-m-thick as corresponding to the *Protoconodontus* Zone correlative with a part of the Upper Cambrian Batyrbaian Stage of the reference sections in the Malyi Karatau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Ordovician
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Cambrian
Upper
Middle
Lower
Batyrbaian
Aksaian
Botomian
Procono-dontus
W. ampli-cava
Epiphyton Beds
4
3
2
1
Alternating siltstone and sandstone with gritstone to conglomerate lenses and interbeds. Trilobites and conodonts Phakelodus tenuis Müll., Protoconodontus sp., and Prosagittodontus cf. dahlmani Müll. in limestone blocks
38
Alternating beds of variegated sandstone and siltstone with shale
55
Alternating beds of variegated sandstone, siltstone, and shale (shale prevails)
51
Alternating beds of variegated siltstone with ocasional sandstone interbeds
5
Rhythmical alternation of lilac and brown–red beds of sandstone, siltstone, and shale
91
Alternating beds of green chlorite shale and gray–green siltstone. Base of siltstone members is sometimes marked by fine-grained sandstone
5
Green chlorite shale
14
Alternating beds of gray sandstone and shale
30
Alternating gray and gray–green beds of chlorite shale and siltstone
25
Dark gray medium-bedded limestone with chert lenses and interbeds. Conodonts Phakelodus tenuis (Müll.), Protoconodontus posterocostatus Müll., Prosagittodontus dahlmani Müll., and trilobites
200
Green– to dark–gray thin-bedded chert. Lenses and interbeds of dark gray to black limestone with conodonts Westergaardodina amplicava Müll, Phakelodus tenuis (Müll)
10–15
Interbiohermal breccia with angular clasts (to 7 m) of algal limestone in the calcareous–tuffaceous matrix. Algae Epiphyton inexpectatum in limestone blocks
80–85
Biohermal algal limestone with obscure bedding. Algae Epiphyton durum, E. induratum, E. frondosum, E. inexpectatum, E. furcatum, E. sp., Girvanella problematica, and Proaulopora glabra; brachiopods Kutorgina sp.
Basaltic tuff and agglomerate tuff conglomerate of the Kursai Formation

**Fig. 3.** The structure of the Karagaily Formation in the Vodopadnyi–Kembriiskii interfluve area (symbols as in Fig. 2).

Range (Dubinina, 1982; Apollonov, *et al.*, 1988). From the same part of the section, P.P. Misyus and A.M. Mambetov sampled trilobites of the "Troedsonian" assemblage: *Lotagnostus hedini* Troed., *L. asiaticus* Troeds., *Charchagia norini* Troeds., *Hedinaspis regularis* Troeds., and *Trinodus suni* Troeds. (determinations by M.K. Apollonov). In addition, these layers yielded inarticulate brachiopod *Paternia* sp. of the uppermost Cambrian (determination by L.E. Popov) and problematics remains, which have not been identified so far.

The total thickness of the member is 225–235 m.

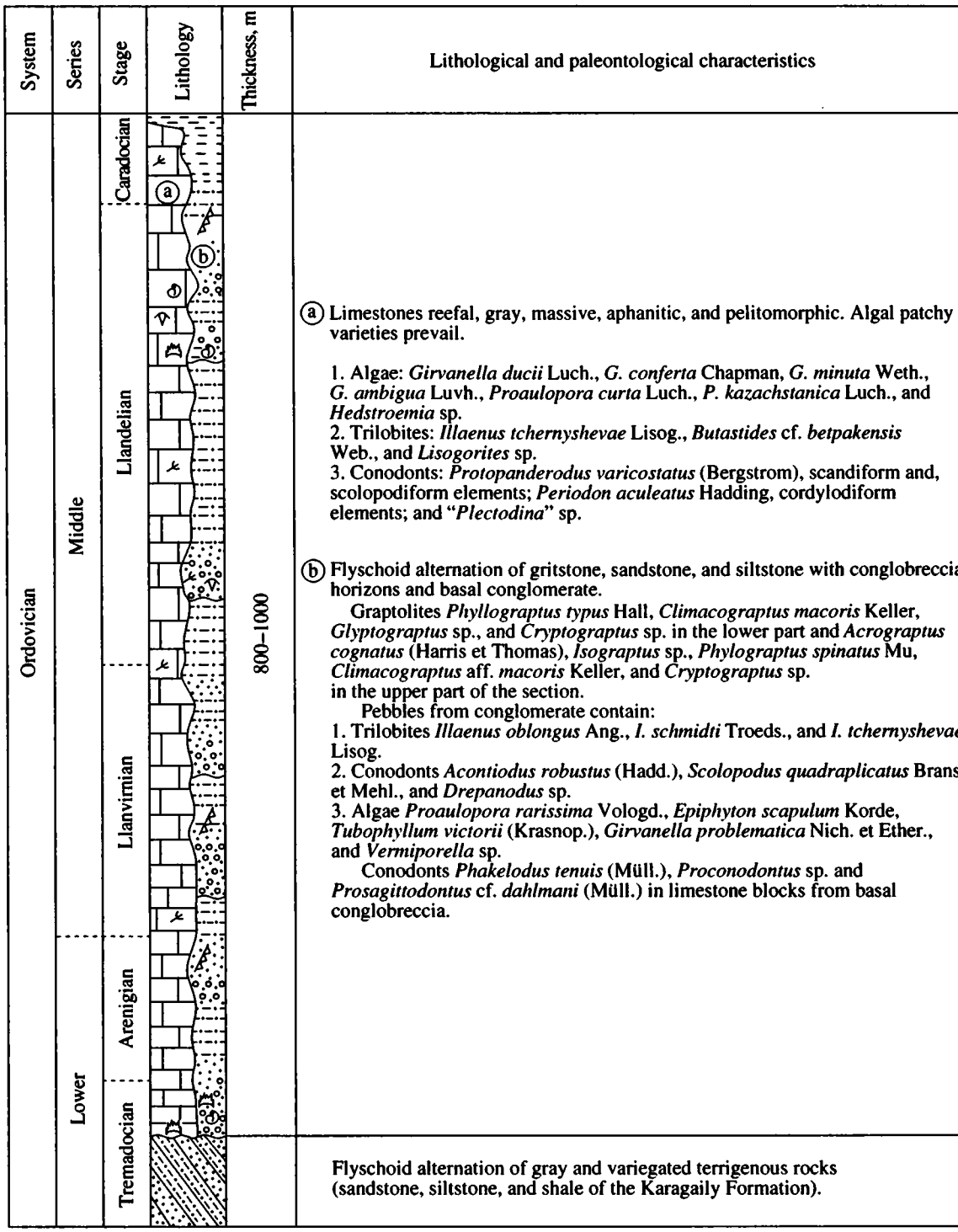
System
Series
Stage
Lithology
Thickness, m
Lithological and paleontological characteristics
Ordovician
Middle
Lower
Caradocian
Llandelian
Llanvirnian
Arenigian
Tremadocian
800–1000
ⓐ Limestones reefal, gray, massive, aphanitic, and pelitomorphic. Algal patchy varieties prevail.
1. Algae: Girvanella ducii Luch., G. conferta Chapman, G. minuta Weth., G. ambigua Luvh., Proaulopora curta Luch., P. kazachstanica Luch., and Hedstroemia sp.
2. Trilobites: Illaenus tchernyshevae Lisog., Butastides cf. betpakensis Web., and Lisogorites sp.
3. Conodonts: Protopanderodus varicostatus (Bergstrom), scandiform and, scolopodiform elements; Periodon aculeatus Hadding, cordylodiform elements; and "Plectodina" sp.
ⓑ Flyschoid alternation of gritstone, sandstone, and siltstone with conglobreccia horizons and basal conglomerate.
Graptolites Phyllograptus typus Hall, Climacograptus macoris Keller, Glyptograptus sp., and Cryptograptus sp. in the lower part and Acrograptus cognatus (Harris et Thomas), Isograptus sp., Phylograptus spinatus Mu, Climacograptus aff. macoris Keller, and Cryptograptus sp. in the upper part of the section.
Pebbles from conglomerate contain:
1. Trilobites Illaenus oblongus Ang., I. schmidti Troeds., and I. tchernyshevae Lisog.
2. Conodonts Acontiodus robustus (Hadd.), Scolopodus quadraplicatus Brans. et Mehl., and Drepanodus sp.
3. Algae Proaulopora rarissima Vologd., Epiphyton scapulum Korde, Tubophyllum victorii (Krasnop.), Girvanella problematica Nich. et Ether., and Vermiporella sp.
Conodonts Phakelodus tenuis (Müll.), Proconodontus sp. and Prosagittodontus cf. dahlmani (Müll.) in limestone blocks from basal conglobreccia.
Flyschoid alternation of gray and variegated terrigenous rocks (sandstone, siltstone, and shale of the Karagaily Formation).

**Fig. 4.** The structure of the Kirtabulga Formation in the Kirta–Karagaily interfluve area (symbols as in Fig. 2).

The upper part of the Karagaily Formation is represented by a terrigenous sequence with the flyschoid rhythmic interbedding. According to the prevalent coloration and composition, it is divided into two members: the lower member (*Member 3*) about 140 m thick and formed by green and gray siltstones and shales with subordinate interbeds of fine-grained sandstone, and the upper variegated member (*Member 4*) about 160 m thick and displaying a greater proportion of sandstone among siltstones and shales.

Our paleontological investigations of the Karagaily Formation resulted in the following.

Abundant algal remains point to the presence of the Lower Cambrian deposits in the section. This stratigraphic interval is represented by biohermal limestones confined to the base of the formation. The topmost biohermal limestones (biohermal chain II) can probably be ascribed to the lowermost Middle Cambrian. We define this part of the section as *Epiphyton* Beds. Organogenic–algal rocks forming a pair of chain-like bioherms correspond, in their morphology, to "biohermal massif" (*Sovremennye* ..., 1990). We suggest to term them as the Karagaily biohermal chain. It may be correlated with a great confidence to the Dedebulak biohermal chain of southern Tien Shan (the Karachytar Ridge), which occupies the similar stratigraphic level.

Top layers of the cherty member are correlated with the *Westergaardodina amplicava* Zone of conodonts Late Cambrian in age. This fact also indicates the normal stratigraphic succession of the Karagaily Formation. The overlying *Protoconodontus* Zone, which is recognized among platy limestones, confirms their correlation with the uppermost Cambrian deposits (the Batyrbaian Stage of Apollonov ). The *Westergaardodina amplicava* and *Protoconodontus* zones mark the reference level for correlation of Cambrian sections of the Aktyuz–Boordy area with those of central Tien Shan. These zones were also recognized in the Sandalash Formation of the Chatkal–Sandalash area of central Tien Shan (Mambetov, 1993).

As previously mentioned, the Karagaily Formation terminates with a horizon of the flyschoid alternation of gray and variegated terrigenous rocks up to 300 m thick. Aside from sponge spicules, we failed to discover other organic remains in the unit. By analogy with other areas, the formation can be attributed to the uppermost Cambrian–Lower Ordovician (Tremadocian). In its position and lithology, this unit is similar to the Tuyuksai Formation of the western Kyrgyz Ridge, Dolonata Formation of the Chon–Kemin area, or their analogues. The apparent total thickness of the Karagaily Formation amounts to 750 m, but usually it is less thick. This depends on the depth of the pre-Ordovician erosion of the core of the synsedimentary Karagaily anticline. At the modern erosion level, this structure represents a one-sided horst–anticline. On the west and east, it is truncated by submeridional faults, the maximum distance between which does not exceed 3.5 km. The occurrence of biohermal limestones composing the base of the Karagaily Formation directly on tuff conglomerates of the upper member of the Kursai Formation can be observed in a large tectonic block located in the interfluve of the Kembriiskii and Vodopadnyi creeks .

*The Kirtabulga Formation.* The Karagaily Formation is overlain with a sharp angular and azimuthal unconformity by the Kirtabulga Formation of the Lower–Middle Ordovician. Its basal boulder–pebbled conglomerates truncate the Karagaily Formation in direction from the northeast to southwest and successively overlie (from the north to the south) the flyschoid, cherty, and limy members of the latter (Fig. 1). The section of the Kirtabulga Formation consists of two different units (Fig. 4). In a greater part of its distribution area, it is composed of monotonous rhythmical alternation of greenish gray and gray gritstone, siltstone, and shale, while, in the remaining area, its characteristic rocks are reefal limestones enclosed into the flysch sequence. The base of some rhythms is marked by conglomerate horizons. Gritstone and sandstone show a low degree of roundness and extremely poor sorting of clastic material. These rocks show a straight graded bedding indicating the normal occurrence of beds. Pebbles and boulders in conglomerates consist of limestone, clayey and siliceous shale, sandstone, and basic and intermediate volcanics, or less commonly, of granitoids, chert, quartzite, and quartz porphyry. Limestone blocks from basal conglomerates yielded remains of conodonts *Phakelodus tenuis* (Müll.), *Protoconodontus* sp., and *Prosagittodontus* cf. *dahlmani* (Müll.) in association with fragments of trilobites characteristic of the uppermost Cambrian. V.I. Goncharova sampled blocks of massive limestone in the upper conglomerate horizons, where they bear remains of Middle Ordovician trilobites *Illaenus oblongus* Ang., *I. schmidti* Troeds., and *I. tchernyshevae* Lis. From the limestone blocks of this part of the section, we sampled abundant remains of Ordovician algae: *Girvanella problematica* Nich. et Ether., *Proaulopora rarissima* Vologd., *Epiphyton scapulum* Korde, and *Tulophyllum victorii* Krasnop. These conglomerates represent a peculiar interbiohermal facies, which encloses, in addition to angular blocks of reefal limestone, the rounded pebbles derived from distant sources; this feature differs these conglomerates from Cambrian interbiohermal breccia. Chemical treatment of limestone blocks and boulders from the upper horizon revealed the conodont remains determined by V.P. Chernyshuk as *Actinodus robustus* (Hadd.), *Scoliopudus quadriplicatus* Brans. et Mehl., and *Drepanodus* sp. of the late Tremadocian–initial Arenigian of the Early Ordovician. Trilobites, crinoids, articulate and inarticulate brachiopods, and algal remains are also known from these deposits. The latter were determined by Luchinina as belonging to *Epiphyton scapulum* Korde, *Tulophyllum victrorii* Krasnop., and green alga *Vermiporella* sp.

Consequently, the erosion affected not only the Cambrian carbonate sequences, but also the Ordovician deposits of the Karagailybulak reef that is exposed among Lower–Middle Ordovician flish deposits of the adjacent Kirtabulga block located to the west.

The interfluve area between the Vodopadnyi Creek and Kirtabulga River comprises a thick monolith of light to dark gray massive limestones and dolomites, which was termed by N.I. Doroshenko and V.I. Kiselev as the Karagailybulak reef ranging in age from the Middle Cambrian to Middle Ordovician. In the explanatory notes to the geological map of scale 1: 500000, the lower boundary of reefal limestones was dated back to

the Early Cambrian (*Stratifitsirovannye* ..., 1982). This erroneous inference resulted from the incorrect localization of Lower Cambrian archaeocyathids and algae sampled by A.G. Vologdin in 1942. After revision of his report, we understood that these remains were sampled from biohermal limestones of the Karagaily Formation (the Vodopadnyi–Kembriiskii interfluve area). This explains the futility of attempts of later researchers to repeat sampling of Cambrian fossils from limestones of the Karagailybulak reef.

In 1951, E.L. Butakova sampled remains of Middle Ordovician trilobites from the upper part of the reefal massif at the northern slope of the Mt. Izvestnyakovaya (altitude of 2645.8 m). In 1957, D.E. Elyutin discovered characteristic Ordovician gastropod *Maclurites* cf. *magnus* Le Sue in the southern part of the reef. We carried out sampling of organic remains over the entire reef beginning from its southerly outcrops. As a result, we were first to discover relatively representative assemblages of the Ordovician algae and Early–Middle Ordovician conodonts. Remains of trilobites, brachiopods, and gastropods were also found at previously known and new sites (Fig. 1).

The oldest Ordovician layers are exposed in the southeastern part of the Karagailybulak reef, near the Vodopadnyi Creek mouth.

As mentioned above, the reefal limestones are overlain in this area by the conglobreccia of the flysch sequence containing algae remnants. Abundant among algal are *Girvanella problematica* Nich. et Ether., *Proaulopora rarissima* Vologd., *Renalcis pectunculum* Korde, and *Epiphyton scapulum* Korde, all of the Ordovician age. Not far away of the indicated site, limestones from the lower part of the reef yielded conodonts determined by Chernyshuk as *Paltodus inconstans* Lind. of the Early Ordovician (Tremadocian–Arenigian) age.

Aphanitic and fine-grained limestones located 20 km south of the northern contact with flysch deposits contain trilobite species determined by Apollonov as *Illaenus tchernyshevae* Lisogor, *Bumastidis* cf. *betpakensis* Web., and *Lisogorites* sp. These fossils are associated with Middle Ordovician algae *Proaulopora curta* Luch. and *Girvanella ambigua* Luch. (determination by Luchinina). Thick-bedded aphanitic limestones, which are exposed 40–45 m to the south of the contact, contain abundant conodonts *Protoponderodus varicostatus* (Bergstrom), scandodiform etl., scolipodiform el., *Perodon aculeatus* Hadding, cordylodiform etl., oistodiform etl., and "*Plectodina?*" sp. (determinations by S.V. Dubinina, Geological Institute, Russian Academy of Sciences). The same rocks yielded trilobite remains, crinoid tabulae, and algae *Proaulopora curta* Luch. and *Girvanella ambigua* Luch. These fossils are characteristic of the Middle Ordovician deposits ranging in age from the uppermost Llanvirnian to Llandeilian and, probably, basal Caradocian (?).

At the left-hand side of the Kirtabulga River 240–250 m southwest of the above site, thick-bedded patchy algal limestones incorporate abundant algae *Girvanella ducii* Luch., *G. conferta* Chapman, *G. minuta* Weth., *Proaulopora kazachstanica* Luch., and *Hedstroemia* sp., as well as ostracodes.

Near the elevation mark of 2352.5 m, massive limestones host remains of algae *Hedstroemia* sp., *Proaulopora curta* Luch., and *Girvanella conferta* Chapman. In Luchinina' s opinion, this algal assemblage can be dated as Middle Ordovician in age. Remains of Ordovician algae *Proaulopora curta* Luch. and *Girvanella ambigua* Luch. were also discovered in the southern slope of Mt. Izvestkovaya, 720–750 m south–southeast of the elevation mark of 2645.8 m.

The synchronous formation of the reef and flysch deposition is also confirmed by other fossils. At the right-hand side of the Kirtabulga River, 600–1000 m west of the elevation mark of 2645.8 m, P.P. Misyus sampled from the flysch sequence the Llanvirnian graptolites *Phyllograptus typus* Hall, *Ph. spinatum* Mu, *Climacograptus macoris* Keller, *Acrograptus cognatus* (Harris et Thomas), *Glyptograptus* sp., and *Isograptus* sp. (determination by D. T. Tsai), synchronous to the fauna from the upper part of the reefal limestones.

## PECULIARITIES IN GEOLOGICAL STRUCTURE OF THE KARAGAILYBULAK REEF

The body in question is triangular in plane. Its southern part is truncated by the Alpine thrust inherited to a certain degree after the ancient tectonic contact between the Cambrian–Ordovician Kastek trough and the Aktyuz–Tegerment massif representing an uplift of the lower Precambrian basement. The eastern and western sides of the triangle are partly complicated by submeridional faults. Northward of the massif, several isolated outcrops of biohermal limestones stretch in the submeridional direction parallel to the flysch bedding. Similar bodies also occur westward of the reef. In Korolev' s opinion, these bodies are not completely isolated but join in the subsurface the main body of the Karagailybulak reef and represent its "satellites."

The western submeridional fault complicating the boundary of the Karagailybulak reef coincides with the boundary between the main reef massif and the biohermal satellites located westward and northward within the Kirtabulga flysch formation. There are three satellites branching off in the western direction. On the eastern side of the triangle, the southern part of the fault passes through the reef. To the west of the fault, the reefal limestones are overlain by conglomerates, which, in turn, are overlain by flysch. In this area, contact between reefal limestones and basal conglomerates is parallel to the northeastward dipping flysch beds. In fact, we deal here with the results of a deeper erosion of the section part transitional between the reefal and flysch deposits in contrast to western outcrops, where the

erosion did not reach a level of the complete junction between the "satellites" and the main reef body. The northern part of the massif represents the rounded top of the triangle, and the upper boundary of the reef massif is conformable here with the bedding of the flysch sequence.

Several outcrops of the biohermal "satellites" slightly isolated from the reef massif and located to the north of it are generally elongated in the submeridional direction parallel to the strike of flysch beds. They form a kind of a body in the anticline core, and conglomerates, which overlap the reef in the southeastern area, adjoin the eastern meridional fault. The westward-going elevation of the structure axis results in nearing the reef summit to the modern erosion surface and, as a consequence, exposes the positive forms of the reef body. Thus, the so-called "satellites" or bioherms do not represent isolated bodies but are closely connected with the massif at deeper levels.

The structure outlined by the "satellites" and overlying flysch beds, including those of conglomerates, is also traced in the Karagaily block, where basal beds of the Lower Ordovician conglomerates and their lower contact with the Cambrian–Ordovician core outline a similar structure. Thus, both the reef and the adjacent flysch beds jointly are the members of the Ordovician structural stage.

Another peculiarity should also be noted. The whole reefal massif is set into the anticline core, whose axis in the flysch deposits stretches in the submeridional direction. The onlap relations, when the carbonate body of biogenic origin is in contact with flysch deposits different in their stratigraphic level, and the joint folding of the carbonate body and enclosing flysch beds suggest the synsedimentary character of the Ordovician structural complex and facial relations between the carbonate platform and flysch.

This conclusion is supported by the fact that the Kirtabulga flysch adjacent to the reef massif encloses graptolites of the Llanvirnian age, and the northern part of the reef bearing the Karakan trilobites is of the same age.

Data on the relationships between flysch and reefal "satellites" show that flanks of the latter have gradual transition from the pure carbonate deposits to a zone of alternating beds of limestone, calcareous conglomerate, and sandstone, documenting, demonstrating a fast replacement of a biohermal to flysch facies. The longitudinal contacts are usually consealed under loose deposits and partly complicated by faults, but when they are exposed, one can observe calcareous conglomerates at the boundary between two facies. Thus, the observed relationships between the rocks indicate the facial replacement, but not the tectonic juxtaposition.

## PRINCIPAL RESULTS AND CONCLUSION

(1) Representative organic remains of the Early Cambrian age were first discovered in the Kursai Formation.

(2) Based on abundant algal remains, the Lower Cambrian age of the Karagaily Formation is substantiated. The topmost layers of the upper (II) biohermal chain can be referred to the basal Middle Cambrian, whereas the conodont zone of the Upper Cambrian (Aksaian Stage) is recognized in the upper layers of the cherty member. This indicates the normal stratigraphic succession of the Karagaily Formation rocks. The recognized conodont zones of *Westergaardodina amplicava* and *Protoconodontus* can be used for the correlation of Cambrian sections of the Aktyuz–Boordy area with the Sandalash Formation of central Tien Shan (Mambetov, 1993) and with reference sections of the Malyi Karatau Ridge (Apollonov *et al.*, 1988).

Flyschoid deposits crowning the Karagaily Formation are barren of organic remains. In the northern slope of the Kastek Ridge, deposits of this stratigraphic level are united into the Kaskara Formation of the Middle–Late Cambrian age (Krasnoborodkin, 1985). In stratigraphic position and lithology, the latter is similar to the Tuyuksai Formation of the western Kyrgyz Ridge, Dolonata Formation of Chon-Kemin, and their analogues.

(3) New localities with abundant organic remains of the Ordovician (Arenigian–Llanvirnian) age were discovered when studying the Karagailybulak reef and flysch beds of the Kirtabulga Formation.

Tremadocian–Arenigian conodonts from limestone blocks of flysch conglomerates occurring in the upper part of the formation probably are derived from the lowermost layers of the reef. The Early Ordovician age of the reef is implied by finds of Tremadocian–Arenigian conodonts discovered *in situ* in its extreme southern outcrops. Thus, the inicial age boundary of the Kirtabulga Formation should be displaced down to the Arenigian level. This is even more so as the basal layers of the Karakiin Formation in the northern slope of the Kastek Ridge yielded, in addition to abundant Karakan trilobites, their forms, which do not exclude the Areinigian age of host deposits (Krasnoborodkin, 1985).

(4) The data obtained indicate the synchronous formation of the reef, its "satellites," and flysch beds.

(5) The normal stratigraphic succession of the Kursai, Karagaily, and Kirtabulga formations is substantiated. This confirms the autochtonous character of the Ordovician deposits in the studied area.

Reviewer S.V. Dubinina

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# Cambrian Reef Formation in Russian Territory

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**Abstract**—The Cambrian reef formation was limited by the Early and Middle Cambrian epochs and culminated in the Lenian–early Amgian time. Reef builders of that time were not diverse, and their agglutinative forms prevailed over ahermatypic and, certainly, hermatypic reef-builders. Accordingly, the buildups had a flattened shape and gentle slopes. Relations between reefs and geotectonic structures of the oceanic and continental types are outlined.

**Key words:** *reefs, Cambrian, archaeocyathians*

## INTRODUCTION

The Cambrian period is characterized by the first appearance of organic framework buildups constructed with the participation of live organisms. It was also the period of the most active reef formation in the earth's history. Cambrian reefs were first described in Russia in the late 1930s–1940s (Predtechenskii, 1937; Dzevanovskii, 1942; Kuznetsov, 1946; Maslov, 1949), though such important reef builders as Archaeocyatha were identified even in the middle of the last century. At first, they were erroneously described as plants, and only later, their animal nature was confirmed (Gekker, 1928; Tol', 1895). The reefs in question are located in the Siberian platform and the Altai–Sayan fold belt. These areas were thoroughly investigated both in outcrops and by means of drilling (Zhuravleva and Zelenov, 1955; Vologdin, 1959; Zhuravleva, 1960, 1966, 1972, 1979; Postnikov and Postnikova, 1964; Zadorozhnaya *et al.*, 1973; Zadorozhnaya, 1974, 1975, 1983; *Geologiya rifovykh ...*, 1979; Astashkin *et al.*, 1984; Kuznetsov and Don, 1984; Nikolaeva *et al.*, 1986; Kuznetsov *et al.*, 1987; and others). Later, more or less detailed descriptions of Cambrian reefs appeared from other provinces of Russia and republics of the former USSR (Brodskaya and Kholodov, 1965; Korolyuk and Sidorov, 1975; Vasil'ev, 1977; Belenitskaya *et al.*, 1990; and others). The data on Cambrian reefs have been successfully used for global paleotectonic and paleogeographic reconstructions (Rozanov, 1984, 1985).

Thus, a great body of materials became available for characterizing the Cambrian buildups, and for defining their specific features and relations with various tectonic structures and units. The terminology used in description, is that recommended by a special reference dictionary (Zhuravleva *et al.*, 1990). The term "reef" was used both in a broad sense (s.l.) to denote all buildups composed of sessile organisms, and in a narrow sense (s.s.) to denote moundlike structures formed close to the sea level and, therefore, exposed to the wave influence responsible for the origin of the frontal reef tail composed of clastic carbonate material that was derived from the reef. Unfortunately, the published descriptions do not always allow the precise detrmination of the type of a reef massif.

### *Geographical Distribution of Reefs in Russia*

Cambrian structures were discovered and almost completely studied in the Siberian Platform for the first time (Fig. 1). The stratigraphic scheme used in this article corresponds to those published in the book *Fanerozoi Sibiri* (1984) and in a special paper on the Lower Cambrian in Siberia (Nikolaeva *et al.*,1986); it also incorporates the latest materials of the collective work by Rozanov *et al.*(1992). Recently, some researchers suggested that the lower boundary of the Cambrian should be placed below the Sunnaginian Horizon (Missarzhevskii, 1989; *III Mezhdunarodnyi ...*, 1990; Shishkin and Zhuravleva, 1990; and others), though not all specialists share this opinion (Khomentovskii and Karlova, 1992; Rozanov *et al.*,1992). When the first skeletal archaeocyathians appeared, their role as builders of framework buildups and bioherms was not significant. At that time, archaeocyathians and algae were able to construct bioherms of a small size. However, the following Kenyadian period was characterized by a wide and intensive bioherm formation.

The reef formation occurred in two different facial–paleogeographic zones: (1) the Anabar–Sinyaya and (2) Turukhansk–Olekma belts. The former represented a rather broad shoal extending from southeast to northwest, from the Aldan shield toward the Anabar uplift, and then it was trending westerly in the sublatitudinal direction. This zone and its westerly continuation separated the East Siberian epicontinental basin, situated father southwest (the Turukhan–Irkutsk–Olekma facies

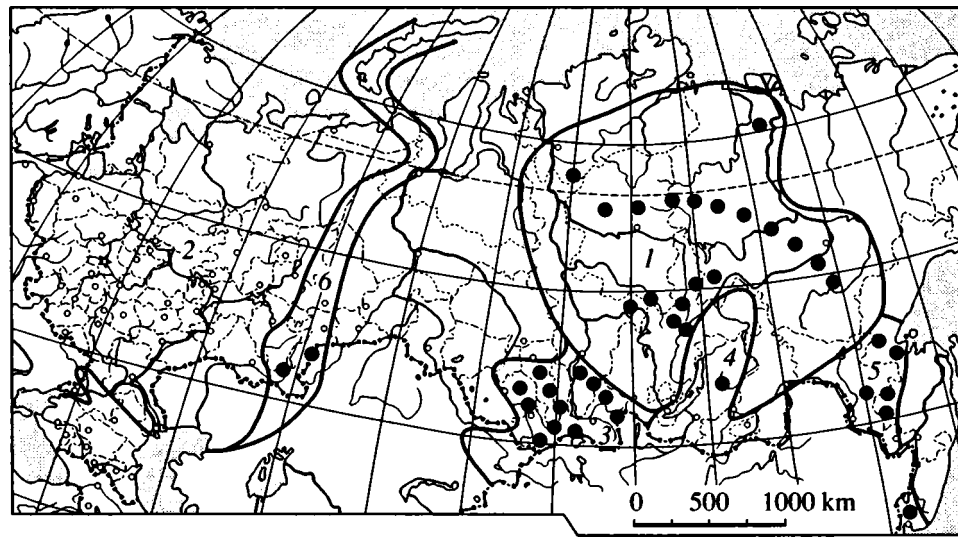


**Fig. 1.** Localities of Cambrian reefs in Russia: (*1*) Siberian Platform, (*2*) East European Platform, (*3*) Altai–Sayan foldbelt, (*4*) Baikal foldbelt, (*5*) Amur–Okhotsk foldbelt, (*6*) foldbelt of the Urals.

belt), from an open sea in the north and northeast (the Yudoma–Olenek facies belt).

Reef buildups of the Anabar–Sinyaya Zone were thoroughly investigated in natural outcrops and, partly, by means of boreholes (Zhuravleva, 1966, 1972, 1979; Luchinina, 1975, 1985; *Geologiya rifovykh* ..., 1979; Astashkin *et al.*, 1984; Nikolaeva *et al.*, 1986; and others). As was mentioned earlier, it was an elongated and rather wide (some 150–200 km) shoal that accumulated carbonate sediments under conditions of normal water salinity. Numerous organic buildups, like biostromes and bioherms, formed within that shoal. In addition, several large massifs (the Oimura, Amga, and others) were distinguished and described there. They are composed of limestones and dolomites. Archaeocyatha, (mostly) blue-green algae *Epiphyton*, (to a lesser extent) *Renalcis* acted as main reef builders. Along with bioherms, various clastic (coarse-clastic including) structures were widespread. There is a disagreement in classification of the latter. In opinion of Zhuravleva, they cannot be regarded as reefs, because some features, (primarily, the absence of the breakwater crest) differ them from younger "real" reefs. She suggests to term these structures as reefoids. Astashkin (1984, pp. 20–25) presented a more complicated model of the structure and development of reefs. The lower part of the Oimura massif about 130–150 m thick and several km wide was dated back to the upper Aldanian (Atdabanian according to the recent stratigraphic scheme of Rozanov *et al.*,1992). In this part, 70–80% of the massif consists of dolomitic algal and archaeocyathian-algal bioherms. An insignificant amount of clastic carbonate material, morphological disorder, and monotonous facies surrounding the massif suggest that it was formed at the stage of the biohermal buildup. The upper part of the massif of the lower Lenian (=Botomian) age is 300–400 m thick and up to 25 km wide. It is composed of algal, archaeocyathian–algal, and diverse clastic carbonates. Brachiopod detritus was encountered as an admixture to remains of reef builders. The massif is situated at the boundary between different facies: at a distance of 1.5–2 km farther north, it is replaced by dark–gray and black limestones, clayey limestones, and combustible shales, while to the southwest it is substituted by shallow-water deposits of the "back-reef shoal." The character of bedding revealed the past morphology of the massif that was elevated over the basin facies by no less than 100–150 m and surrounded by gently dipping slopes. The description of the Amga massif (the Amgian Stage of the Middle Cambrian) is much the same; numerous trilobites were recorded there (Astashkin, 1979, 1984).

Thus, the clear morphology and facies asymmetry suggest that, at least, from the second half of the Early Cambrian, an asymmetrical reef system was generated in this area and evolved in the Amgian period of the Middle Cambrian, when its progradation towards the deep sea was in progress. Buildups of the later time (the Mayan Stage) are represented exclusively by biostrome massifs (for instance, the Tankhai–Ust Mil complex).

Cambrian rocks of the Turukhansk–Irkutsk–Olekma facies belt are overlapped by younger deposits, and organic buildups were explored here by drilling (Postnikov and Postnikova, 1964; Grachevskii *et al.*, 1976; Kalinkina *et al.*, 1981; Kuznetsov and Don, 1984; Kuznetsov *et al.*, 1987, 1992; Kuznetsov and Postnikova, 1988). The Cambrian section of the belt is represented by intercalated carbonate and saliferous members. The reef formation was associated here with transgression epochs, when organic buildups appeared in carbonate members. Buildups of the Osa Horizon (uppermost parts of the Tommotian Stage) have been

best studied, as they are of the productive type and well characterized by core samples. They are located in the two slope areas of two vast shoals corresponding to the large tectonic structures: the Nep–Botuoba and Baikit anteclises. Sometimes, they occur on their dome parts. Initially, the reef builders were represented by *Renalcis*, later associated with *Epiphyton* clusters, and then with *Givanella*, *Marcovella*, stromatolites, *Catagrafia*, and oncolites. Archaeocyatha species are also present among them, but they are more scarce here than in the Anabar–Sinyaya zone (Kuznetsov *et al.*, 1987).

The buildups are rather large in area (40–100 km$^2$), but their thickness is no more than 100–120 m and slopes are extremely gentle. The buildups exhibit an asymmetrical geometry: their slopes facing depressions are steep (about 1°–2°), while more gentle slopes look toward the dome part of the anteclise (0.2°–0.3°). Simultaneously, their elevation above the seaward bottom (90–150 m) is much greater than above the rear shoal zones (usually a few meters and less than 10–15 m). By their type, the buildups in this zone most likely correspond to biohermal mounds. The Altai–Sayan foldbelt reveals abundant and diverse reef buildups, which have been studied in detail (Zhuravleva *et al.*, 1962; Zadorozhnaya *et al.*, 1973; Zadorozhnaya, 1974, 1975, 1983). This zone is rich in biostromes, bioherms, and biohermal mounds up to 1000 m in thickness, and even in real reefs, some 2000 m thick. Among their reef builders, there are *Epiphyton*, *Renalcis*, *Razumovskaya*, *Khabarovia*, *Sayania*, *Girvanella*, and other taxa, Archaeocyatha included. Organic buildups are associated here with different sedimentary formations: terrigenous, volcanic–terrigenous, terrigenous–carbonate, and carbonate. The age of buildups in different places is dated by the Lower Cambrian stages; in some cases, they were also formed in the early Middle Cambrian (the Tortash reef massif at the slope of the East Sayan anticlinorium). Most buildups, particularly large structures, are confined to tectonic uplifts of different kinds and to their slopes. In troughs, they are partly substituted by deepwater deposits.

The comprehensively studied Tortash complex (Zadorozhnaya, 1983) is of special interest for this analysis and represents a great isolated lens-shaped ridge, about 2 km thick. It is Atdabanian–early Middle Cambrian in age. The complex mostly consists of archaeocyathian–algal limestones, carbonate breccias, and a subordinate amount of platy limestones. The rocks are to a great extent recrystallized and, locally, completely converted to dolomites. Several lithologic–facial zones are identified in the ridge: (1) massive algal-frame and subframe limestones, (2) laminated platy detritic limestones, (3) carbonate breccias, consisting of fragments of calcareous algae, etc. Some may not accept my straightforward actualistic interpretation of these zones as corresponding to reef plateau, intrareef lagoon, and other elements. However, facies differentiation, lateral facies association, abundant breccias of framework limestones (about 60–70% in some zones) confirm that it was a real reef ridge with clearly identified paleogeomorphology. Zadorozhnaya believes that the reef formation proper was activated twice: in the Bazaikh–Kameshkovo (Atdabanian) time and in the terminal Early–Middle Cambrian period. The break in the reef formation during the Botomian period may simply reflect the lack of necessary information or facies migration, etc. According to Rozanov (personal communication), the reef formation lasted in this case until the late Atdabanian–Botomian.

Two more points are significant: (1) a relatively small amount of framework limestones composed exclusively of *Epiphyton* remains with negligible amount of Archaeocyatha; (2) paleoecological differentiation of biota and its occurrence mode. Within the zone of the reef edge and most active hydrodynamics, only *Epiphyton* species could act as reef builders; their thalluses were compact, nearly spherical in shape with shortened branches, and they formed dense feltlike structures. In more quiet back-reef zones of the same reef complex, branching *Epiphyton* thalluses had an elongated form and created the large-cell lattice. In these conditions, the biota composition was more diverse and included such forms as *Razumovskaya*, *Proalupora*, *Sayania*, *Khabarovia*, and others, which, however, did not produce frame structures being dispersed as isolated lumps supplying the carbonate material. Archaeocyathians were also locally abundant, but their remains occur in allochthonous burials. Therefore, frame structures proper are absent in this reef complex, while blue-green algae acted during its development as filtrating, agglutinating, and stabilizing organisms.

In other regions of Russia and former soviet republics, the reef buildups have been studied to a lesser extent. Within the Hercynian fold system of the Urals, Lower Cambrian reef buildups have been described in the Sakmara (South Urals) and East Uralian zones (Korolyuk and Sidorov, 1975). Biostromes and biohermal mounds up to 100–150 m wide and 50–100 m thick, but usually smaller, occur here among volcanic–terrigenous deposits. Isolated buildups are joined to form submeridional chains extending, in some cases, for several dozens of kilometers. Among the reef builders, there are *Epiphyton*, *Renalcis*, *Khabarovia*, oncolites, and Archaeocyatha. Large buildups are characterized by zoning: they have a oolite–oncolite basement, a massive core, and brecciated edges with abundant archaeocyathians.

Cambrian reef buildups have been also described as constituents of the Atdabanian carbonate-volcanogenic formation in the Baikal foldbelt extending along the southern border of the Siberian platform. Reef formation proceeded here in two stages separated by epoch of intense volcanism of acid and intermediate composition (Vasil'ev, 1977). Reefs at the slopes of uplifts were situated 200 m above the fore-reef basin, and had an asymmetrical structure in cross section. *Epiphyton*, *Razumovskaya*, and diverse Archaeocyatha were the

main reef builders. Planktonic forms *Amganella* and *Proalupora* served as a source of carbonate material.

Small biostromes, bioherms, and biohermal mounds, usually several meters thick and occasionally ten times thicker, were identified to the southeast of the Siberian platform, in the Amur–Okhotsk fold system (Belenitskaya *et al.*, 1990). They occur as members of the Atdabanian, Botomian, and lowermost Toyonian sequences composed of terrigenous, volcanic–sedimentary, and terrigenous–carbonate deposits. Along with the common reef builders of the Cambrian time (blue-green algae and archaeocyathians), remains of inarticulate brachiopods and trilobites were recovered from reef deposits.

Isolated algal-archaeocyathian mounds of a broad stratigraphic range, some 800 m thick, occur amid terrigenous rocks in the western piedmonts of the Trans Ili Alatau Ridge (Hercynides of northern Tien Shan) outside of Russia, in republics of the former USSR. The age of the lower part of the mounds is Lower Cambrian, while its upper part is Middle Ordovician. The detailed description have been reported for Cambrian buildups in the Karachatur Range of Kyrgyzstan (Teslenko *et al.*, 1983). Bioherm formation started in the Early Cambrian (the early Sanashtykgolian time in the Altai–Sayan region), and further, after a considerable hiatus in sedimentation, it continued in the Amgian period of the Middle Cambrian. The initial buildup at the base of all massifs is represented by taphostrome, in other words, by a biostrome grading upward into the bioherm that preserves some biostrome parts. Among organic remains, diverse archaeocyathians, algae, hyoliths, brachiopods, gastropods, and monoplacophorans were detected. Individual bioherms are 6–8 m thick. They compose a sedimentary body 35–40 m thick, whose shape resembles an atoll-like semicircle 1.2 km long and situated around the past volcanic center. The bioherm ridge is surrounded by spilite-keratophyre lavas; breccias; and keratophyre tuffobreccias, tuffs, and tuffites intercalated with black siliceous dolomites.

### *Tectonic and Paleotectonic Position of Reefs*

The considered patterns of reef distribution show that reefs are associated with various types of tectonic structures. In the ancient platform, they are interrelated with carbonate and saliferous–carbonate deposits and situated on flexural inflections (the Anabar–Sinyaya facies belt) and on gently dipping slopes of platform anteclises. Outside of these tectonic elements, buildups are scarce and thin, and mostly represented by simple bioherms. In foldbelts, reefs (s.l.) are developed in inner paleoridges of different type and scope, mainly at the edge of slopes. They are associated with a wide range of formations: terrigenous, volcanic–terrigenous, and terrigenous–carbonate.

In the genetic aspect, it would be more interesting to review the position of reefs using the model plate tectonic reconstructions, although it is difficult to select the most credible and suitable model from the many available. The most popular are global reconstructions by Scotese and McKerrow (1990), McKerrow *et al.* (1992), and the somewhat updated maps by Golonka *et al.* (1995). Rozanov (1985, 1992) also suggested paleogeographically logical reconstructions for the Early Cambrian. However, all reconstructions, seemingly quite credible, are based on worldwide data and can hardly be used in this work. They depict large-scale structures, mostly continental plates, while microcontinents, stable blocks in past oceans, and zones of subduction and volcanism are missing. In other words, these maps are inappropriate for detecting the paleotectonic position of reefs in the Altai–Sayan, Urals–Baikal, and Amur–Okhotsk regions. Reconstructions made by Zonenshain *et al.* (1990) for the territory of the former USSR appear to be most suitable for this consideration, despite some inaccuracies concerning the Cambrian epochs noted by Rozanov. The positions of major plates (Siberia, Europe, Gondwana, and its elements) are very similar on all maps of paleotectonic reconstructions. Some latitudinal differences are insignificant (5°–10°) and do not go outside the accuracy of interpretation, although the longitudinal discrepancies are somewhat more appreciable. Disagreement in the positions of Arctida and certain Gondwanan blocks are not important for this work, because these tectonic elements are either situated outside the study area, or do not contain reefs and, therefore, cannot influence the subsequent conclusions. An advantage of the maps by Zonenshain *et al.* is their large scale and detalization. Accordingly, they offer the possibility to analyze the paleotectonic position of reefs in fold areas, though not always with the expected accuracy.

In general, the past distribution of reefs is quite regular (Fig. 2). According to the paleotectonic scheme, the Anabar–Sinyaya reef system had a sublatitudinal trend and was situated approximately within 8–12°S. Inner reefs of the Nep–Botuoba and Baikit anteclises were formed in the subequatorial zone. I should note that the intense salt accumulation was concentrated in the same zone.

The reefs of fold areas are related to the submeridional Paleoasian ocean. Many of them (reefs of northern Tien Shan and, partly, of the Altai–Sayan zone) were formed within the Chingiz–Salair–Minusinsk–Tuva island arc system, where they were associated with volcanosedimentary formations. Other reefs were dispersed along the margins of ancient microcontinents (the Tuva–Mongolian, Barguzin, and others) and associated with carbonate and terrigenous–carbonate formations. The Sakmara zone (South Urals) was, apparently, related to the subduction zone of the East European platform and Uraltau, the latter being a part of the Trans-Urals block (microcontinent). This geodynamic position was responsible for the association of reefs with the volcanic–terrigenous formations.

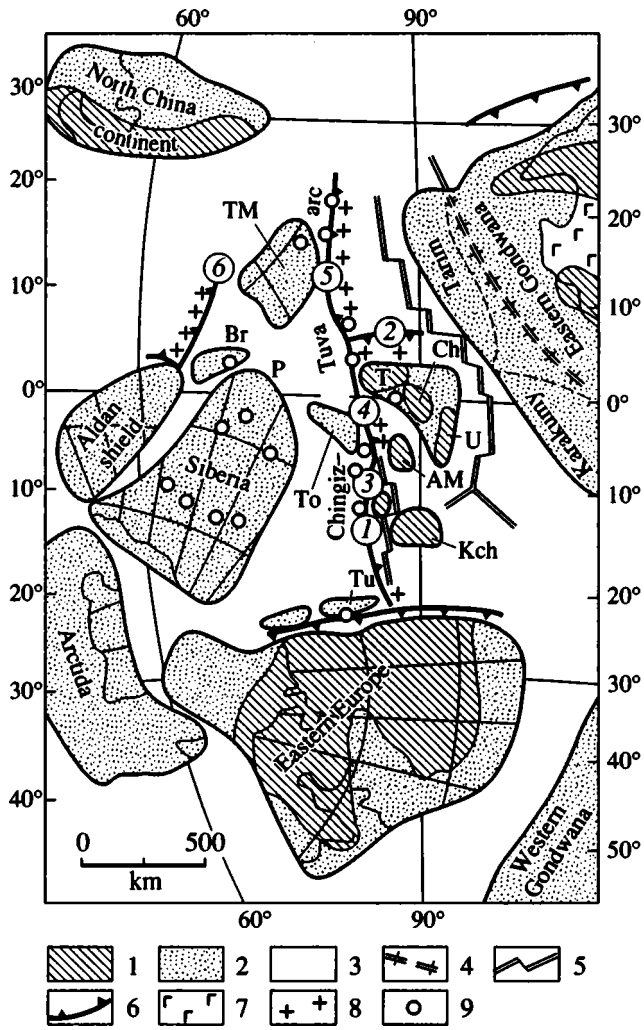


**Fig. 2.** Position of reefs in the paleotectonic reconstruction of the Early Cambrian (540 Ma) after Zonenshain *et al.* (1990): (1) land, (2) shallow seas, (3) oceanic basins, (4) continental rifts, (5) spreading axes, (6) subduction zones, (7) traps, (8) calc-alkaline volcanism of island arcs and active continental margins, (9) reefs. Precambrian and other massifs and large-scale allogenic blocks: (AM) Atasu-Mointin, (Br) Barguzin, (Tu) Transurals, (Kch) Kokchetav, (T) North Tien Shan, (To) Tomsk, (TM) Tuva–Mongolia, (U) Ulytaus, (Ch) Chuisk, (P) Paleoasian ocean. Volcanic arcs (figures in circles): (*1*) Chintinskaya; (*2*) North Tyan Shan; (*3*) Salair; (*4*) Minusinsk; (*5*) Tuva; (*6*) Vitim.

The position of the reefs in the Amur–Okhotsk system, and particularly in the Khingan–Bureya massif and surrounding foldbelt, is less clear, as "the available materials are insufficient for the correct reconstruction of the pre-Devonian history" of the area (Zonenshain *et al.*, 1990, Book 1, p. 306). Two hypotheses are suggested: (1) reefs were located close to the Tuva–Mongolian massif, or (2) in the north China continent (ibid.). The former version based on the similarity of sedimentary facies seems more credible and is shown on the presented map (Fig. 2).

Thus, the adopted paleotectonic reconstruction suggests that all Cambrian reef buildups were concentrated in the equatorial zone restricted by 15°–20°N and 20°–23°S.

### *Some Characteristic Features of the Cambrian Reef Formation*

The available materials on the Cambrian reef formation indicate that it started in the Early Cambrian by the appearance of bioherms and bioherm mounds. Reef formation attained its maximum in the Lenian and, probably, in the early Amgian periods, when reef breakwaters were generated. This geological process declined in the Mayan epoch and virtually ceased in the Late Cambrian (Fig. 3).

Biota of the Cambrian buildups is rather peculiar. It includes *Renalcis*, *Epiphyton*, *Girvanella*, *Marcovella*, *Khabarovia*, *Razumovskaya*, stromatolites, and various problematics, like *Catagraphia*, oncolites, and Archaeocyatha. Among accompanying organisms there are trilobites, inarticulate brachiopods, *Porifera*, planktonic *Amganella* and *Proalupora*. Despite such a primitive composition, a definite succession of communities and rock types can be detected in the section: this is a kind of prototype for the succession typical of younger reefs.

The pioneer populations of blue–green algae are represented by diverse microphytolites often lacking the clearly shaped fragments and calcitization. They produced agglutinant useful for trapping fine material (clay particles including) and the facilitated cementation of sediments. This led to the formation of accumulative forms weakly expressed in the sea floor relief. These buildups were mostly composed of clayey micritic dolomites. Later, the more intense illumination stimulated generation of more complex biocenoses consisting of *Renalcis*, which later were supplemented by crusty *Marcovella*, branching epiphytes, and Archaeocyatha. This community composed the bulk of the construction consisting of polyphytic limestones, often dolomitized. By the end of the reef formation, under conditions of more active hydrodynamics and higher salinity in the reef shoal, branching forms were again replaced by cementing crusty forms, and further by microphytolites.

Another specific feature of the biota is its functional composition. The real hermatypic builders of reefs were almost missing. Their function can be performed, and, in part, was performed, by Archaeocyatha exclusively, though their role in this aspect was limited. In most cases, the abundance of archaeocyathids was insufficient for the creation of the reef framework. Moreover, in many cases, Archaeocyatha remains occur in the redeposited state, rather than in the autochthonous position. Under these conditions, they commonly served as a substratum for blue-green algae and cyanobacteria. Bushy forms of *Epiphyton*, because of their size and friability, could hardly produce a solid framework as well; most likely, they acted as cementators of sedimentary material (and naturally as its producers). Organisms-cementators apparently played the dominant role in forming the Cambrian buildups,

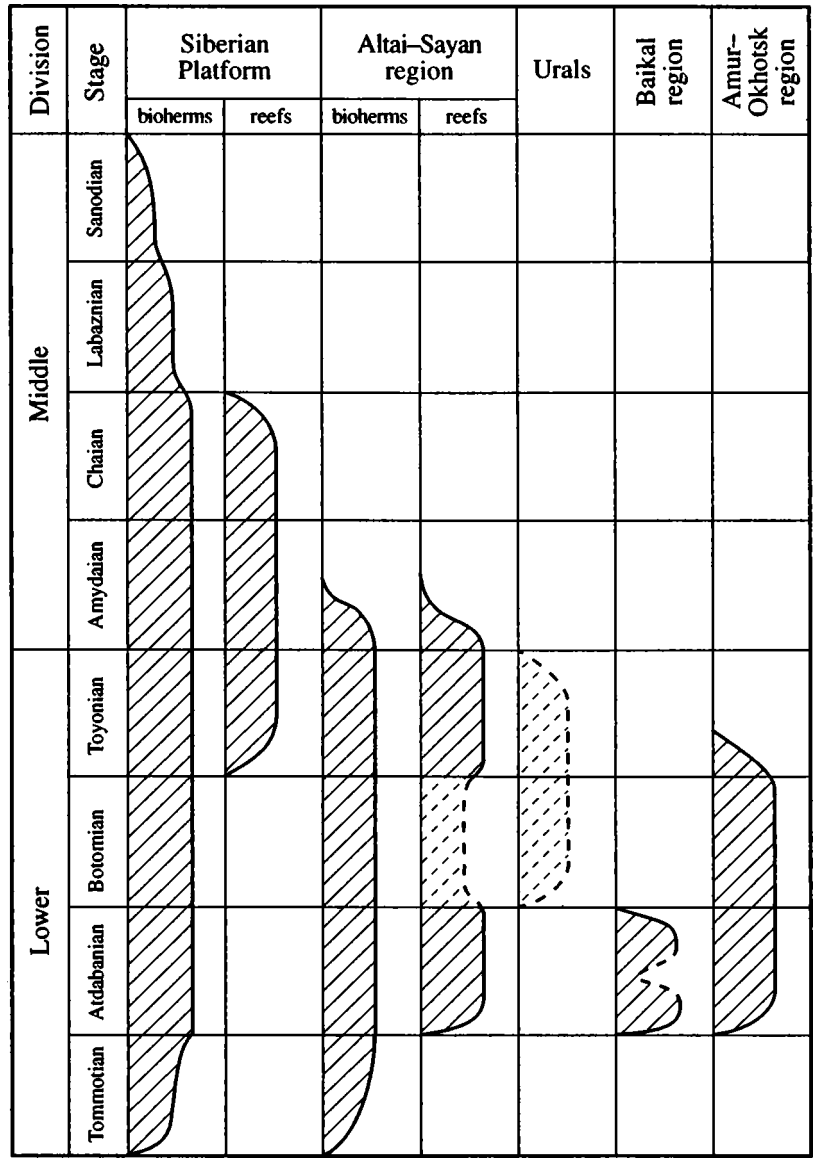


**Fig. 3.** Stratigraphic distribution of reefs in Cambrian deposits of certain Russian regions.

whereas agglutinative and, in particular, hermatypic reef builders played a minor role. This evidently determined the peculiar morphology of buildups as we will see below. Despite the insignificant abundance, the very appearance of vertically arranged and bushy epiphytes and Archaeocyatha species greatly contributed to the formation of buildups with clearly exhibited morphology and even to the reef creation, though the mechanism of this function remains unclear.

These frameless reefs, usually termed as "mud mounds," are also known in younger deposits. Their features and formation conditions have been actively discussed in several works (Bourroque *et al.*, 1983; Flügel, 1987; Somerville *et al.*, 1992; Flajs and Hüssner, 1993; Leinfelder *et al.*, 1994).

Finally, the third specific feature of biota, particularly of archaeocyathids, is, as established by Zhuravleva (1972, Fig. 37), the earlier appearance of many genera and species, their greater diversity, and the later disappearance from the organic buildups, if we compare them with the coeval host deposits. These features are characteristic of not only Cambrian buildups, but all reefs, in general. For instance, the earlier appearance and wider diversity of stromatoporoids in Ordovician and Silurian reef facies was reported by Klaaman and Nestor (1968) from Estonia. A uniform biota of Cambrian reefs accounts for a poor diversity of their rock types, predominantly represented by micrograined polyphytic and, rarely, archaeocyathid limestones and dolomites.

The Cambrian reef buildups (s.l.) apparently include all the types: common organic buildups, such as biostromes, bioherms, or their assemblages in a form of large biostrome and bioherm massifs and, finally, the reefs proper. The occurrence of large and thick bioherm massifs is quite common. The reefs proper (s.s.) usually form asymmetrical reef systems, whereas atolls and atoll-like reefs have not been described yet. These buildups commonly occupy vast areas and have extremely gentle slopes responsible for the flattened form of buildups. This is a consequence of the specific nature of reef-building organisms mainly represented by cementing ahermatypic forms, whereas hermatypic builders played an insignificant role. These features distinguish the Cambrian buildups from the Upper Paleozoic and, certainly, from the Mesozoic–Cenozoic well-outlined reefs with steep slopes, though they do not exclude a reef affinity of Cambrian structures. Moreover, James and Geldsetzer (1988) were quite right in stating that "every ancient reef is unique," and, therefore, the straightforward actualist approach to identifying types of ancient reefs (s.s.) is hardly appropriate.

It took a geological age to convert the bioherms into the reef massifs (s.s.), because, in the Altai–Sayan region, the latter already appeared in the Atdabanian Stage. In the platform, the reefs emerged later, in the Lenian and Amgian stages. The problem of the earlier reef (s.s.) generation in the mobile zone, as compared with the platform, is of particular interest. The facts suggest that Archaeocyatha did not migrate into the Altai–Sayan region from the Siberian platform until the early Atdabanian (Debrenne *et al.*, 1989; Rozanov, 1992). The transition from bioherm massifs to reefs (s.s.) might have been a reaction to the wide Atdabanian transgression and concurrent sea-level rise (Kirshvink *et al.*, 1984). To secure the position in the ecological niches colonized earlier in the shallow water zone, the reef builders should increase the rate of biogenic carbonate accumulation and vertical growth that finally resulted in appearance of reefs. From this viewpoint, it is understandable why reefs appeared earlier in mobile areas: the rapid subsidence in the latter required the adequately accelerated growth of buildups, i.e., their transformation into reefs.

## CONCLUSION

The reef formation (s.l.) in a form of archaeocyathid–algal buildups started in the Early Cambrian and virtually ceased in the late Middle Cambrian. It was at its maximum in the Lenian–early Amgian time, when reefs (s.s.) were formed in addition to the bioherm massifs. The latter appeared in fold belts earlier than in platforms.

The Cambrian buildups are characterized by a poor taxonomic diversity of reef builders (blue-green algae and, to a lesser extent, archaeocyathids), among which the ahermatypic organisms were much more abundant than hermatypic forms. This specified the peculiar morphology of buildups with their flattened forms and gently dipping slopes. The appearance of skeletal organisms such as Archaeocyatha and upright branching epiphytes notably promote the formations of buildups with clear morphology elevated above the sea bottom.

All Cambrian buildups were localized in the subequatorial belt, approximately between 15–20°N and 20–23°S, where they were interrelated with definite tectonic structures and, accordingly, with particular geodynamic regimes, each characterized by its own association of sedimentary formations. The reefs of ancient platforms are associated with carbonate and saliferous–carbonate formations. Reef buildups of ancient blocks (microcontinents) or, more precisely, of their slopes, are interrelated with carbonate and terrigenous–carbonate formations. And, finally, reefs of subduction zones in the front of island-arc volcanic structures intimately associate with volcanic–sedimentary deposits, the siliceous clayey formations including.

The Cambrian reef formation was an important stage of this process, however, not perspective. It proceeded during the Early and initial Middle Cambrian, producing archaeocyathid-algal buildups, and then ceased in the Late Cambrian. After a long period, the reef formation resumed in the late Ordovician on a fundamentally new biological basis. Despite all intricate deviations, the major trend of reef formation was related to vital functions of algae and coelenterates (Ravikovich and Zhuravleva, 1975; Termier and Termier, 1975; Kuznetsov, 1983, 1986; Kuznetsov, 1990; and others). However, certain lateral and dead branches occasionally appeared in this trend, including those associated with cyathiform organisms homological to archaeocyathids, such as sponges of the late Jurassic, rudists of the middle Cretaceous, etc. (Kuznetsov, 1990).

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# Transgressive–Regressive Cycles in Evolution of the Arctic Ocean during the Late Cenozoic

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**Abstract**—Three main stages related to major transgressive–regressive cycles of the development are recognized in the Late Cenozoic history of the Arctic Ocean: the Miocene–early Pliocene (regressive), late Pliocene–Eopleistocene (transgressive), and Pleistocene–Holocene (regressive). Transgressions and regressions were mainly induced by tectonics and related to global oscillatory movements, which included rhythms of the lesser order. They controlled the degree of the Arctic Ocean isolation from adjacent oceanic basins and changes in the ecological system particularly pronounced in the shelf area. The transgressions were accompanied by a wide migration of marine fauna, whereas the regressive stages expanded areas of cryogenic sedimentation.

**Key words:** *Arctic Ocean, transgression, regression, Late Cenozoic, correlation, fauna, cryogenesis*

The Arctic Ocean occupies a particular place among other oceanic basins. It is located in the circumpolar part of the Northern Hemisphere and isolated, to a great extent, from the adjacent Atlantic and, particularly, Pacific oceans. At the same time, it represents a sea basin connecting these two oceans. During some evolutionary stages of the Arctic Ocean, its connection with both oceans almost completely ceased, whereas, on the contrary, it was quite vivid in other periods and widened the faunal and floral interchange between the North Pacific and North Atlantic.

The Arctic Ocean is the smallest of all oceanic basins, occupying only $1 \times 10^6$ km$^2$, i.e., about 3% of the World Ocean, and, simultaneously, it is the youngest. According to the concept of mobilism, it was formed as a deep water basin at the Mesozoic–Cenozoic boundary time (*Geologicheskoe stroenie* ..., 1984). Our paper is focused on the Late Cenozoic (Miocene–Pleistocene) period of its geological history.

As is known, the Miocene was a period of global land elevation and sea floor subsidence. By the end of the Miocene, the relative sea level was by 350–500 m lower than at the beginning of the Paleogene as a result of both the increase in dimensions and height of continents and the growth in the volume of oceanic basins. This is thought to be responsible for the global temperature fall by 2–3°C (Markov and Velichko, 1967). All the events were accompanied by the cardinal reorganization of the structure in the Arctic and many other regions of the world. Some morphological structures oriented sublatitudinally, for instance, the circumpolar continental shoal appear in the north of Eurasia and North America that occupied the area of the modern and past Arctic shelf and adjacent to deeper parts of the Arctic continental slope. Plains located in this region accumulated mainly continental subaquatic (alluvial, lacustrine, and others) deposits.

The Miocene continental deposits are recorded in the northern areas of West and East Siberia (Danilov and Smirnova, 1976; Polyakova and Danilov, 1989; Fradkina *et al.*, 1989). They are represented by alluvial kaolinite sands with interbeds and pellets of kaolinic clay and lenses of lignite, as well as by lacustrine micaceous silt, sandy loam, loam, and clay. In the areas of northeastern Siberia adjacent to the modern coastal zone of the East Siberian and Chukchi seas, the Miocene time was characterized by the accumulation of lagoonal clayey–silty sediments with lignite and lenses of lignitic peat (Danilov, 1980). Lagoonal deposits are distributed strictly along the coastal zone of the mentioned sea basins and occur 50–70 m below the modern sea level. Consequently, there are grounds to suggest that the extreme northeastern Siberia was occupied in the Miocene by a shelf zone, which was close in its parameters to the modern one but formed at lower relative sea level. Similar deposits were characteristic of the Beaufort Formation in the Arctic coast of North America and Canadian Arctic Archipelago, where it mostly consists of continental deposits (Crag and Files, 1965; Matthews, 1976, 1989). The formation distinctly bounds the southern areas of maritime plains and is dated as the Miocene in age, at least in its lower part, where its spores–pollen spectra are represented up to 50% by pollen of broad-leaved plants. In the Main Island (Canadian Arctic Archipelago, 80°N), continen-

tal deposits alternate with marine sediments (Matthews, 1987).

Miocene climatic conditions in the Arctic are ambiguously understood and seem to be significantly variable during the time period from 25 to 5–7 Ma ago. There are various paleofloral data on the Arctic islands, particularly, on the northeastern Siberia, where areas near the coast of Arctic seas were characterized in the fist half of the Miocene with the mixed coniferous–small-leaved forests incorporating an admixture of relatively thermophilic plants such as beech, oak, chestnut, maple, walnut, and others. Many researchers believe that the mid-Miocene time was characterized by the preponderant distribution of broad-leaved forests typical of the Miocene climatic optimum (Fradkina *et al.*, 1989).

The development of thermophilic vegetation in the circumpolar coastal areas of the Arctic Ocean set one to assume an easy connection of the ocean with the adjacent Atlantic and Pacific basins, although the direct evidence of this is missing. The wide distribution of Miocene diatoms probably redeposited in sediments of the northern Chukchi Peninsula and occuring at about 100 m above the sea level (Danilov, 1980) should be mentioned as an indirect evidence. The active influx of warm waters of the Atlantic and Pacific origin into the Arctic basin explains also its warm environments in the early and middle Miocene.

Abundant palynological data obtained, in particular, those for northeastern Asia (Fradkina *et al.*, 1989) suggest a significant cooling in the Arctic regions during the late Miocene and glaciation in the southern Alaska mountains, where moraines alternate with lava flows, whose age is determined by the K-Ar method. According to this dating, glacial events occurred 12–10, 9–8, and 7–5 Ma ago (Armentrout *et al.*, 1978).

Recently, Polyakova (1997) obtained new data on Miocene diatom assemblages from coastal lowlands of the northern Chukchi Peninsula. The first encountered index-species of marine diatoms allowed her to correlate their assemblages with the zonal stratigraphic scale of the North Pacific region. As a result, two stages of influx of Pacific waters into the Arctic Ocean were established, and these events were dated back to 11.5–10.5 (or 8.9–8.4) and 6.6–5.1 Ma ago. In addition, there are indications that the Arctic Ocean was covered with ice as early as the initial late Miocene. This is evident from the presence of cryiophilic species typical of the cryopelagic biocoenoses of marine ice and from the glacial–neritic diatom assemblage recovered from the relevant deposits. The occurrence of cryophilic forms and the arctic–boreal type of the diatom assemblage directly indicates the cold water conditions in the ice-covered basin located in the coastal plains of the northern Chukchi region during the late Miocene.

According to palynological data (Zyryanov *et al.*, 1992), the Chukchi Peninsula coast of the late Miocene sea basin was occupied by the dark coniferous taiga of the northern type, which later (at the end of the Miocene) was replaced by the vegetation of the antecedent hypoarctic zone.

The presence of the ice-rafted material in sediments dated back to the beginning of the late Miocene, i. e., to approximately 10 Ma ago (Thomas and Thiede, 1991), is also recorded in the Atlantic sector of the Arctic Ocean (the Norwegian–Greenland sedimentation basin). The first maximum of the ice-rafted material concentration at 9.5–7.0 Ma ago apparently indicates the opening of the Fram Strait and initiation of the East Greenland Current, which carried waters from the Arctic basin into the North Atlantic.

If we accept that the Arctic Ocean within the subarctic zone was covered with ice as early as in the initial late Miocene, then it is reasonable to assume that ice appeared in the Arctic sector, particularly in its central part, most likely at the end of the middle Miocene. According to Norwegian researchers, the middle–late Miocene boundary interval marks the initiation of permanent connections between the Arctic and North Atlantic oceans and the inflow of Arctic waters into the latter (Kristoffersen, 1990). In other words, the water mass circulation similar to the modern one commenced at that time in the Arctic and Subarctic regions, and there are grounds to presume that the formation of the ice cover initiated at the end of the middle Miocene in the Arctic seas and in the ocean as a whole. This is evident from the occurrence of ice-rafted rounded pebbles in middle Miocene lagoonal loam in the northern Chukchi Peninsula coast (Danilov, 1980). In the subarctic zone of the Pacific coast of Alaska (Middleton Is.), layers of glacial and glacial–marine sediments are recorded in the 1181-m-thick section of marine deposits (the Yakataga Formation) 12.7–13.36 Ma old, i. e., correlated with the uppermost middle Miocene (Miller 1953; Gladenkov, 1988). Matthews (1976) suggested an appearance of tundra landscapes and permafrost in the Canadian Arctic in the middle Miocene. Data on marine faunal assemblages allow an assumption that the local glaciation in the northeastern coast of the Gulf of Alaska commenced as early as the beginning of the early Miocene, i.e., about 20 Ma ago (Gladenkov, 1988).

The ultimate structural reorganization of coastal areas of northern continental margins and the formation of an integral, mainly sublatitudinal curcumpolar morphological structure of the Arctic shelf, including its past zone, occurred during the Miocene–Pliocene boundary time (*Geologicheskoe stroenie* ..., 1984; Danilov, 1985; Afanas'ev *et al.*, 1988). This means that the further development of this integral structure of the coastal shelf zone corresponds to the Pliocene–Quaternary (Recent) stage of the earth's geological history. Beginning from the terminal Miocene–initial Pliocene, the geologic–geomorphologic evolution of this morphological structure occurred synchronously in general and was related to the large-scale oscillatory tectonic

movements of the global character (Danilov, 1985; Afanas'ev *et al.*, 1988) responsible for the alternation of transgressive and regressive stages in its evolution. Transgressions and regressions were the principal events of the Recent geological history of the Arctic Ocean and adjacent lowlands of northern Eurasia and North America. They had a decisive effect on the paleogeography and, in particular, on the climate of the entire Northern Hemisphere (Danilov, 1982). During the transgressive stages, connections between the Arctic basin and global ocean widened and the inflow of relatively warm Pacific and, particularly, Atlantic waters intensified. On the contrary, the isolation degree of the Arctic Ocean increased during the regressive stages and resulted in its cooling, which considerably influenced the adjacent continents.

Three stages related to major long-term oscillatory movements are distinguished in the recent history of the Arctic Ocean, mainly of its shelf zone. The first stage, corresponding to the terminal Miocene–initial Pliocene, is marked by the general elevation of the northern continental margins of Eurasia and North America, by draining of the Arctic shelf with the development of denudation processes and the active incision of paleorivers into the peneplain, and also by the deepening of oceanic basins. The second stage, corresponding to the second half of the Pliocene (3.3–1.8 Ma) and Eopleistocene (1.8–0.8 Ma), was a time of preponderant subsidence and wide development of transgressions over the vast plains of northern Eurasia. Seawater flooded the northern areas of the European part of Russia and ingressed into the north of the Yana–Kolyma lowland, coastal plains of the Chukchi Peninsula, and other Arctic and Subarctic continental areas. Finally, the third stage (Pleistocene within the range of 0.7 m.y. and Holocene) was generally characterized by elevations in the Arctic shelf and areas adjacent to it on the south.

Against the background of general elevation related to the oscillatory rhythm of the high order, there were oscillations of a lower second order, which resulted in a short-term subsidence and intermittent transgression. This is reflected in the formation of terrace-like relief and marine terraces along the Arctic sea coasts located now at the altitude of 200–220 to 2–4 m (Danilov, 1978, 1985). Whereas these events were synchronous in general and induced by the oscillatory tectonic movements of the global scale probably accompanied by eustatic sea-level fluctuations, there were also the regional differences related to peculiarities in the tectonic structure of particular areas. The scale and duration of transgressions and regressions were different in different areas. Regional peculiarities were pronounced during the entire Recent geological history of the Arctic.

The first stage, corresponding, as mentioned, to the terminal late Miocene (the Messinian sea-level fall) and initial Pliocene, was a time of major regression manifested, to a greatest extent, in the continental margin of Northern Eurasia. The regression in various structural–tectonic regions was, however, different in intensity (Danilov, 1978). In the southern coast of the Pechora Sea, the base of ancient river valleys is now buried at the altitude of –350 to –400 m. In the Yana–Kolyma Lowland, the bottom of ancient valleys is located 150–200 m below the modern sea level. Finally, in the Chukchi Sea coast, the incision level of the early Pliocene valleys is 80–100 m below the latter. These ancient relief forms are typical of the modern Arctic shelf.

Comparing the basal depth of the late Miocene–early Pliocene valleys with the depths of adjacent seas we may conclude that the Arctic shelf virtually did not exist at that time; this area was almost completely represented by land (Fig. 1). This increased the isolation degree of the Arctic Ocean from the adjacent Pacific (complete cessation of the connection) and Atlantic (substantial decrease in the water exchange). The isolation of the circumpolar basin naturally affected its hydrology and thermal regime. The inflow of warm waters into the Arctic Ocean sharply decreased and inevitably resulted in the cooling of its water and the appearance of the permanent ice cover.

Unambiguous records of the regular influx of the coarse-grained ice-rafted material into the central deep part of the Arctic Ocean are characteristic of the Miocene–Pliocene boundary period. The youngest age of deposits with the ice-rafted material is 5.6 Ma; the date was estimated on the basis of sedimentation rates and paleomagnetic data (Clark and Morris, 1984).

The study on Upper Cenozoic deposits in the northern lowland coast of the Chukchi Peninsula shows that frozen grounds and high taiga-type forests of larch, spruce, burch, and willow appeared here as early as in the early Pliocene (Danilov, 1980). The lower Pliocene alluvial shingle with buried tree trunks and ground pseudomorphs after veins of thawed ice occur now at the depth of more than 50 m below the zero isobath. This means that the greatest part of the Chukchi Sea did not exist at that time. Another area located to the north of the drained shelf zone was occupied by the ice-covered sea basin, because the permafrost indicators in the flat coastal land suggests the presence of a stable ice cover in the adjacent sea (Danilov, 1982, 1985).

The second stage, comprising almost the entire Pliocene–Eopleistocene period, corresponds to the descending branch of the major oscillatory cycle. This stage is mainly marked by the transgressive development of the continental margin, which surrounded the deep (bathyal to abyssal part of the Arctic Ocean). It was a time when the sea flooded spacious areas of maritime plains, which turned into the vast Arctic paleoshelf substantially exceeding in dimensions the modern one. The transgression intensity varied in different areas depending on their structural–tectonic position. In the western segment of Eurasia, including the Pechora, West, and North Siberian lowlands

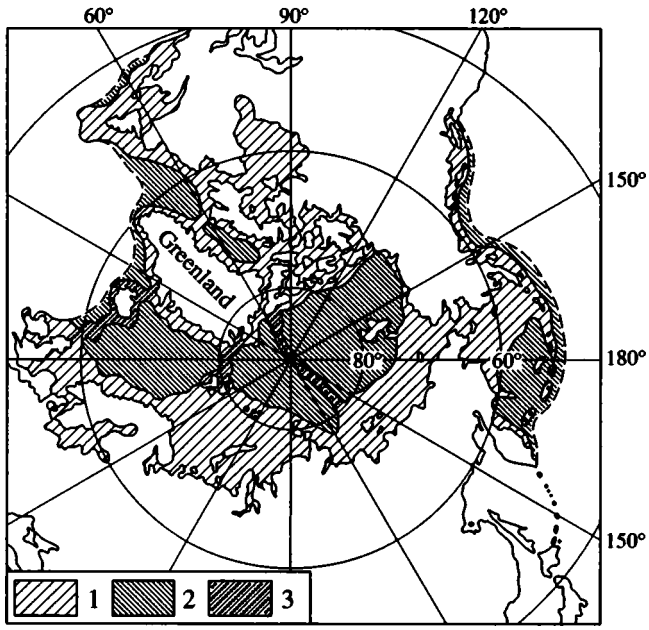


**Fig. 1.** The Arctic Ocean and adjacent oceanic basins during the stage of a major regression at the Miocene–Pliocene boundary: (1) dried shelf areas; (2) deep oceanic areas; (3) shallow oceanic areas (barriers for the Atlantic waters).

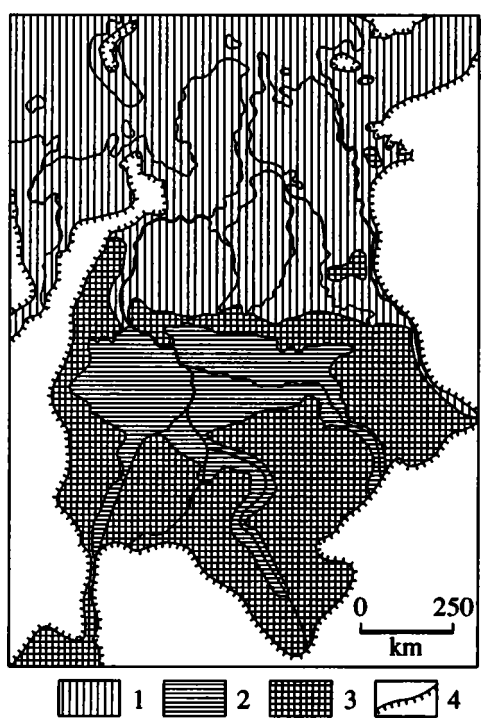


**Fig. 2.** Paleogeographic scheme illustrating the maximum development of the Late Cenozoic (late Pliocene–Eopleistocene) transgression in northeastern areas of the European part of Russia and in West Siberia (after Danilov, 1982).
(1) sea basin; (2) impounded fresh-water basin; (3) shoals, islands, and maritime lowland plains; (4) boundaries of land.

together with shelves of the Barents and Kara seas, transgression of this stage were most spacious reaching in the south the northern ridges in the European part of Russia and in West Siberia (Fig. 2). During the entire transgressive stage, the sea level stand was close to the modern one. Some partial regressions and draining in the Arctic shelf are recorded in sections by coastal–marine, alluvial, and lacustrine sandy deposits of shingle and clay occurring at the altitude close to the modern position of the sea level or slightly above it (Danilov, 1978). Nonetheless, continental deposits, which would occur significantly below the modern level of the Arctic Ocean and could indicate its regression episodes are unknown for the time period in question. The subsidence in the stable plains of the Pechora lowland and in northern areas of West Siberia progressed down to the depth of 200–250 m during the transgression maximum, as is evident from the occurrence level of marine deposits. In the folded mountains, the subsidence and subsequent elevation were of a greater amplitude. For instance, marine deposits occur at the altitude of 375 m above the sea level in the Pai-Khoi Ridge, 420 m in the Polar Urals, and of 320 m in the Putorana piedmonts.

The transgressive stage was marked by the formation of main sequences of marine deposits in accumulative plains of northern areas of European Russia and West Siberia, as well as in the shelf zones of the Barents and Kara seas. These deposits compose the vertical section up to 300–400 m thick (100–150 m in average) and are exposed beginning from the base of ancient pre-Pliocene and early Pliocene valleys to summits of moderns watersheds. Coastal–marine, alluvial, and lacustrine facies corresponding to partial regressions of the Arctic Ocean are distinctly subordinate in the main Cenozoic sequences exposed in the accumulative relief of watersheds. Preponderant facies are represented by loam and clay peculiar in lithology: they are poorly sorted, lacking distinct bedding, and contain ice-rafted material (gravel, pebbles, boulders). In other words, they are of the diamicton appearance. Due to this feature, some of their horizons are often attributed to glacial moraine deposits. Nonetheless, loam and clay in the northern areas of entire West Siberia and in the Pechora lowland enclose remains of marine organisms: foraminifers, ostracodes, mollusks, and crustaceans; in some sections, they host marine malacofauna of the buried banks (Danilov, 1978; and others). The analysis of occurrences of this fauna clearly indicates that it was buried *in situ*, because the thin-walled shells are frequently covered with epidermis, open valves are filled with the host sediment and joined in the clamp, and so on. The surface of some boulders bear the attachment traces of barnacles. All these features directly testify to the accumulation of marine faunal remains and boulders of loam and clay in the cold-water sea basin, where the coarse clastic material was transported by floating ice (marine fast ice and icebergs).

In addition to faunal remains, the marine genesis of loam and clay is evident from the composition of dispersed authigenic minerals: phosphate, iron sulfides (vivianite, hydrotroilite, pyrite), and carbonates (calcite, aragonite, and, less commonly, dolomite). In the areas of local accumulations of these minerals, concretions carbonate (glendonites), pyrite–hydrotroilite, and vivianite were formed (Danilov, 1978; and others). Combination of marine faunal remains with peculiar occurrence features and assemblage of authigenic minerals are characteristic of marine sediments and indicate the glacial–marine genesis of enclosing deposits.

The Upper Cenozoic sequence of glacial–marine deposits composing the accumulative relief of watersheds between the coastal plains of western North Eurasia is characterized by the rhythmical structure. It is divided into three cyclic members, which are termed as the Kolvin, Padimei, and Rogov formations in the Pechora lowland (Vollosovich, 1966; Zarkhidze, 1970). In northern West Siberia, they correspond to the Turukhan (Polui), Mess (Kazym), and Sanchugov (Salekhard) formations (Danilov, 1978). The rhythmic patterns of the main sequence of Recent deposits are related to oscillatory tectonic movements of the second order, which occurred against the background of descending movements of the major tectonic cycle responsible for the generals subsidence of the territory. The second-order oscillatory movements were accompanied by brief partial sea regressions from the continental margin of North Eurasia (Afanas'ev *et al.*, 1988).

Slow subsidence processes in the spacious plains of northeastern Siberia during the period in question were completely compensated by the continental sedimentation. Therefore, these areas are characterized by the development of alluvial, alluvial–deltaic, and lacustrine swampy deposits of the relevant age. Nonetheless, the brackish-water and marine facies of a limited distribution are also present among the latter. They are developed a little wider on the coastal plains of the Chukchi Sea (Danilov, 1980).

The vast transgression of the Arctic Ocean onto the maritime continental lowlands resulted in two consequences. First, the connection with adjacent oceans widened and the influx of warm Atlantic and, to a lesser extent, Pacific waters increased and tempered the local water masses. Second, the cold Arctic water could enter the southerly sea basins and result in their cooling. It is believed that the exchange of molluscan fauna between the Pacific, Arctic, and Atlantic oceans occurred at this stage, in the time periods of about 4.38, 3.6–3.2, and 2.5–1.9 Ma (Gladenkov, 1994). In this very time (2.5 Ma ago), the first cooling episode (pre-Tiglian) is recorded in the recent history of the Netherlands, where several cooling episodes (pre-Tiglian, Eburonian, Menapian) are recognized in the time interval of 2.5 to 0.7 Ma, i.e. in the late Pliocene–Eopleistocene (Haesaerts, 1984). During the period in question, northern (boreal) species of marine fauna migrated into the Mediterranean. It is known that some boreal species, for instance, *Arctica* (*Cyprina*) *islandica* L., whose modern distribution area is limited in the south by the Bay of Biscay (44.5°N), occur in the Calabrian deposits 1.8–0.8 Ma old. It is believed that its displacement by 8.5° (about 1000 km) to the south down to the latitude of 36°N may indicate the lowering of surface water temperature in the Mediterranean approximately by 5°C (Markov and Velichko, 1967). It is also probable that there were several phases of expansion of northern (boreal) fauna to the south in the time period of 3–4 to 0.8 Ma.

According to data of some researchers (Vollosovich, 1966; Zarkhidze, 1970; Yakhimovich *et al.*, 1973; Danilov, 1985; and others), one of these expansion events was related to the Akchagylian transgression of the Caspian Sea (3.3–1.8 Ma ago), when, as they believe, the Arctic and Caspian basins were connected via valleys of the Pechora and Severnaya Dvina rivers, on the one hand, and Kama and Volga rivers, on the other. The idea that the Arctic and Caspian transgressions of the Pliocene time episodically ran together in the European part of Russia was conjectured by A.P. Karpinskii. It is known that the Akchagylian marine fauna is alien to the Miocene and Pliocene deposits of southern Russia. It immigrated only once into the Caspian Sea after the accumulation of a fossil-barren productive sequence and then disappeared as suddenly as appeared.

All the aforesaid confirms that the Arctic Ocean was frequently connected with the adjacent oceans and with some of the interior sea basins.

The third stage in the Arctic Ocean evolution, corresponding to the Pleistocene (in its minimum scope) and Holocene, was mostly marked by the elevation of the Arctic shelf and adjacent continental areas during the ascending branch of the major oscillatory tectonic cycle. Smaller, relatively short-period oscillatory movements of the second order were responsible for the brief Arctic transgressions into maritime plains and also for the intervening regressions that resulted in a partial draining of the shelf. As a result, a series of consistent relief forms and marine terraces, whose height regularly diminishes seaward becoming younger in the same direction, were formed on the North Eurasian plains (Danilov, 1978, 1985; Afanas'ev *et al.*, 1988; and others).

Regional differences were pronounced at this stage of the development of North Eurasian continental margin as well. In the western segment of the territory, the amplitude of oscillatory rhythms of the second order complicating the main rhythm was at the maximum and caused a wide development of high morphological steps and well-manifested marine terraces. The absolute height of these morphological steps is identical in the Pechora Lowland and in the northern areas of West Siberia. The step-like pattern of the relief was noted by

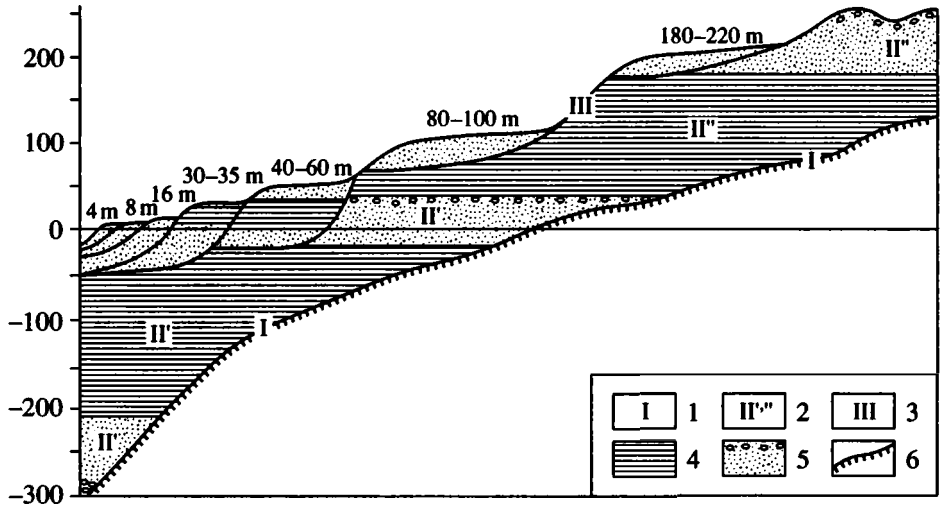


**Fig. 3.** The principal scheme demonstrating main stages in the development of the Arctic and Eurasian (western segment) shelves in the Late Cenozoic (after Danilov, 1985).

(1–3) Stages: (1) deep erosional differentiation of the territory and formation of the late Miocene–early Pliocene buried valleys: (2) formation of main upper Pliocene–Eopleistocene sequences composing accumulative plains with substages (denoted by the accent mark) related to transgressive–regressive cycles of the second order; (3) formation of the watershed relief of accumulative plains, step-like morphological levels, and marine terraces; (4) deep water deposits (clay, silt) reflecting the maximum phases of the sea basin development; (5) shallow-water deposits (sand, gravel) reflecting transgressive–regressive phases of the sea basin development; (6) eroded top surface of the pre-Pliocene deposits.

many researchers who studied the geomorphology of these regions (P.P. Generalov, I.D. Danilov, I.A. Kuzin, T.A. Matveeva, N.V. Reinin, V.N. Saks, P.N. Safronov, V.T. Trofimov, N.G. Chochia, and others). Despite the minor disagrement in estimates, the distinguished morphological levels have altitudes of 4–6, 8–10, 12–16, 25–30, 80–100, 110–120, 150–180, and 200–220 m. They are manifested in the highest accumulative watersheds, whose absolute height is about 240–250 m. In the plains of northeastern Siberia, the general elevation and oscillatory movements of the Pleistocene time produced the step-like patterns of the accumulative relief and a series of terraces. Marine sediments of that time compose the coastal terraces with the altitudes of 2–3, 4–6, 10–12, 20–25, 40, 60, 80, and 100 m.

Figure 3 demonstrates the principal stages in the Late Cenozoic evolution of the North Eurasian continental margin exemplified in the western segment (Pechora and West Siberian plains) of the region. The first stage (I) corresponding to the late Miocene–early Pliocene regression of the Arctic Ocean was a time of active incision and regional erosion. The second stage of the late Pliocene–Eopleistocene transgression (II) was marked by the deposition of sedimentary sequences composing the accumulative relief of coastal plains, the watershed areas inclusive. The figure also shows two transgressive–regressive cycles (II' and II"), which represent substages of the second stage. Facies of the transgression maximum (deep-water fine-grained sediments) and regressive–transgressive facies (shallow-water coastal sand and gravel) are recognized among the deposits of these cycles. Finally, the third stage (III) corresponds to the development of the continental margin in the Pleistocene (in its classic, minimal scope) divisible into 4–6 or 7 and more glacial and interglacial phases. This stage resulted in the formation of a series of step-like relief forms and marine terraces lowering toward the Arctic Ocean; the lower of these terraces are accumulative and the upper ones are of the compound type.

It is interesting to note that a similar structure is peculiar of coastal plains of the Netherlands (Haesaerts, 1984), which are mostly composed of the upper Pliocene–Eopleistocene marine deposits 2.5 to 0.7–0.8 Ma old according to the isotopic dates. The younger Pleistocene deposits form a series of riverine and marine terraces. It seems that the upper Cenozoic deposits of the Baltic region are characterized by a similar structure and formation history (Afanas'ev *et al.*, 1988). This type of coastal lowlands with the accumulative cover of marine deposits was attributed by K.K. Markov to the West Siberian or Dutch type, in contrast to the Mediterranean coastal zones characterized by a series of terraces with decreasing height and age (Markov and Velichko, 1967). Nonetheless, terraces and terrace-like relief forms lowering toward the ocean and simultaneously becoming younger are also characteristic, as shown, of the coastal accumulative plains (Fig. 3). It is important to emphasize that the absolute height of these relief levels of the accumulative plains is not less than the average height of Mediterranean terraces (Fig. 4). The highest Calabrian terraces of the Mediterranean region (typical altitude about 180 m) are most likely Eopleistocene (1.8–0.8 Ma) in

age. The marine accumulative terraces located in the tectonically stable lowland coasts are of the same or greater altitude and, consequently, should be older rather than younger. In any case, one can confidently state that the formation period of the marine terraces in the northern accumulative plains was not less than the range of the Pleistocene epoch. This means that the main marine sequences of the maritime plains (northern West Siberian, Pechora, and Dutch lowlands) are not younger than the Eopleistocene.

Another important conclusion was formulated by F. Zeiner in 1959: inasmuch as the lower marine terraces are younger in age, the sea level steadily lowered during the Pleistocene in response to the ocean bottom subsidence, i.e., in the eustatic mode. Nonetheless, it is not inconceivable that the subsidence of oceanic bottom was combined, as in the Miocene, with a uniform elevation of continents (the ascending branch of the major tectonic cycle or the third stage). The glacioeustatic sea-level fluctuations probably also contributed to the formation of coastal marine terraces, but their role was insignificant.

Despite the extremely low temperatures during the stages of regression and climate cooling, the on-land glaciation could develop only in areas with a sufficient precipitation. Areas with such a regime were Scandinavia, Polar and North Urals, mountain areas of Putorana, Byrranga, northeastern Siberia, Alaska and northeastern America. The intensity of glaciation in the Arctic islands, including Greenland and the Canadian Arctic Archipelago, was hardly greater than now. Glaciation in Siberia was also insignificant, because the cooling alone could not result in formation of the ice cover. According to A.I. Voeikov, an outstanding Russian climatologist, the climate in this region is sufficiently cold even nowadays, but the ice formation requires an increase in precipitation, particularly during the winter season, and this process could be stimulated most likely by the transgression of the Arctic Ocean into the maritime lowlands, for instance, in West Siberia and the maritime plains of the Chukchi Peninsula.

Under modern climatic conditions, the on-land glaciation in the northern hemisphere occupies an area of about $2 \times 10^6$ km$^2$. An area occupied by the subsurface glaciation (beyond the on-land glaciation areas) amounts, according to calculations by I.Ya. Baranov and R. Blacke, to about $20 \times 10^6$ km$^2$, and the latter is 10 times greater than the former. The average ice-covered area in the Arctic Ocean is about $9 \times 10^6$ km$^2$, and the sea-to-land glaciation ratio is 4.5 : 1.0, whereas the proportion between the sea and surface glaciation is 0.45 : 1.0. In other words, the subsurface glaciation distinctly prevails over others in the northern (continental) hemisphere. This situation was hardly different in general during the cryogenic Pleistocene epochs. In particular, the subsurface and related sea glaciations were preponderant during the last late Pleistocene epoch of the severe continental climate, and the on-land glaciation was of a lesser significance.

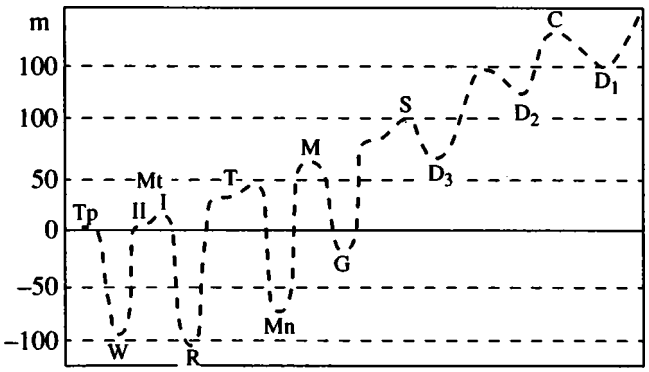


**Fig. 4.** Curve demonstrating the relative sea-level fluctuations based on the results of the study of Mediterranean terraces (the Depere–Zeiner series). High (transgressive) levels: (C) Calabrian, (S) Sicilian, (M) Milazzan, (T) Thyrenian, (Mt) Monastyrian (I and II), Tp (Flandrian Tapes). Low (regressive) levels: ($D_1$, $D_2$, $D_3$) Danubian, (G) Günz, (Mn) Mindel, (R) Riss, (W) Würm.

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# The Comparative Study of the Cyclicity and Rhythmicity in Precambrian Sequences

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**Abstract**—Results of the study of the cyclicity and rhythmicity in Precambrian sequences, i.e., of periodicity in the development of sedimentation basins, magmatism, folding, and metamorphism, are discussed. Cycles and rhythms in the formation of Precambrian rocks are compared with similar Phanerozoic cycles and rhythms. The original cause and interrelated pulsatory development of external and internal geospheres during the entire history of the earth are considered. It is shown that principles of Phanerozoic stratigraphy can be applied in the Precambrian stratigraphy and stratigraphic subdivision of particular Precambrian sections may be performed as thoroughly and reliably as that of Phanerozoic sections. It is emphasized that the level of detailedness in stratigraphic subdivision of the Precambrian and Phanerozoic sections corresponds to the state of knowledge about the Precambrian and Phanerozoic evolution of the earth. The further improvement of a general model of periodicity in geological evolution is assumed to be dependent on the study of development stages of the Precambrian sedimentary basins.

**Key words:** *Precambrian, stratigraphy, hierarchy of geological events, cyclorhythmology, pulsation of geospheres, metrology*

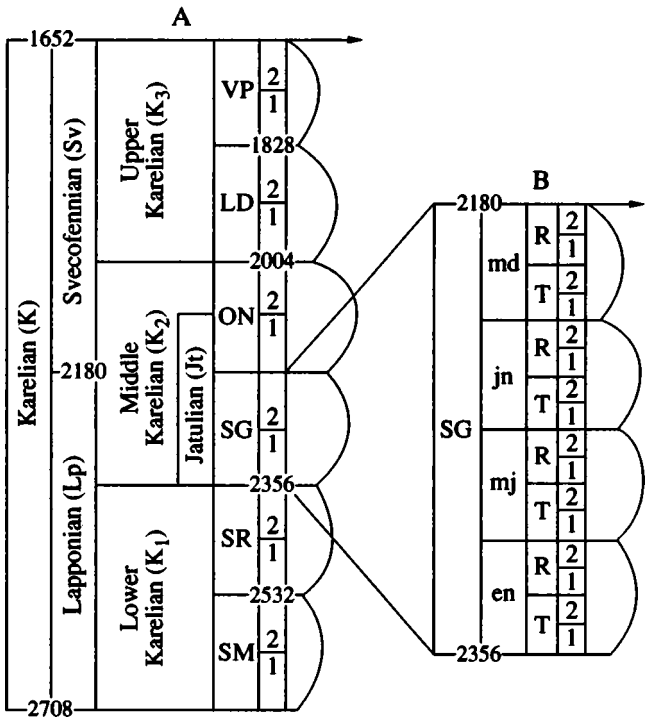
The knowledge of a general periodicity in the earth's evolution considerably depends on the improvement of the standard stratigraphic scale of the Precambrian spanning 7/8 of the entire geological history. The competing approaches that are used now in the Precambrian stratigraphy do not help the solution of this problem (Menner, 1980; Semikhatov, 1991, 1993, 1994; Bekker, 1995; and others). In this regard, principally new possibilities are suggested by the comparative study of stages in the development of sedimentation basins, magmatism, tectonic deformation, and metamorphism (Negrutsa and Negrutsa, 1988; Negrutsa, 1994a, 1994b). Speaking about the stages, we are discriminating between cyclic (directional–reciprocal) and rhythmic (unidirectional) processes.

## PRIMARY PREREQUISITES

In connection with the prospecting of gold–uranium conglomerates in the Precambrian sequences of Karelia and the Kola Peninsula in the second half of the 1950s and first half of the 1960s (Negrutsa, 1990), Russian geologists faced the necessity of compiling advancing facial–paleogeographic maps for the particular time slices. In turn, this required the elaboration of detailed local stratigraphic scales for the subdivision of sedimentary–volcanogenic conglomerate-hosting sequences. Initially, all the potential uranium-bearing deposits were subjected to study (Negrutsa, 1992), but later, the main attention was concentrated on Proterozoic successions. For a long time, these successions represented a model object for the construction of the scales using methods of genetic lithology and correlation of heterofacial sequences barren of organic remains (Negrutsa, 1963; Bezdenezhnykh *et al.*, 1966; Negrutsa and Negrutsa, 1967).

Even the first step of investigations showed that the knowledge of stages and space–time regularities in changes of facial–paleogeographic settings of sedimentation and volcanism is extremely important for elaborating the detailed Precambian stratigraphy (Negrutsa, 1967, 1971). Applied to Karelia, the necessary degree of details and reliability of facial and paleogeographic reconstructions were provided by a large volume of drilling, and geological prognoses were effectively tested by detailed prospecting. It was principally important that all the observations, from description of exposure (drill cores) to general interpretation, were performed on the genetic base and the correlation of sections in the particular tectonic structures was carried out by means of mapping the stratigraphic markers and boundaries of principal units. On the one hand, this substantiated the detailed correlation of local stratigraphic sections and resulted in the separate construction of integral stratigraphic schemes for every structure. On the other hand, we were able to receive knowledge of the spatial regularities in facial, paleogeographic, structural–metamorphic, and geochemical changes.

Such a complex approach provided the unique, for the Precambrian, data. First, the correlation of local




Schematic structure of the main stratigraphic unit of the interregional scale (A) and regional stratigraphic unit (B).

The curve shows waves of transgressions (direction of arrow) and regressions. Vertical columns demonstrate stratigraphic units and their ranks: stratigraphic unit of the second rank—Karelian (K); stratigraphic units of the third rank—Lapponian (LP), and Svecofennian (Sv); stratigraphic units of the fourth rank—Lower ($K_1$), Middle ($K_2$) and Upper ($K_3$) Karelian; stratigraphic units of the fifth rank—Sumian (SM), Sariolian (SR), Segozerian (SG), Onegian (ON), Ladogian=Kalevian (LD), and Vepsian (VP); stratigraphic units of the sixth rank—lower and upper parts of above-listed units of the fifth rank indicated as $Sm_1$, $SM_2$, $SR_1$, $SR_2$, etc.; stratigraphic units of the seventh rank—Eningilampian (en), Majmyarvian (Mj), Jangozerian (jn), Medvezhiegorian (md); stratigraphic unit of the eighth rank—transgressive (T) and regressive (R) rhythms of cycles of the seventh rank; stratigraphic units of the ninth rank— elementary cycles ($T_1$, $T_2$) $R_1$, $R_2$, etc. Numbers in the columns denotes the adopted isotopic age of boundaries between stratigraphic units.

stratigraphic schemes was performed on the facial–paleogeographic and paleogeodynamic basis, a system of chronologically interchangeable features important for correlation of heterofacial sections was thus established, and the first model of event–stratigraphic Precambrian scheme of the Baltic shield was elaborated (Negrutsa, 1963, 1979, 1994b, 1995). Second, the relative rank of Precambrian stratigraphic units was established and the hierarchy of these units, regardless their age, was defined on the basis of comparative geohistorical and chronometric evaluation (Negrutsa, 1984, 1988, 1992). Third, the general genetic cause and the resonance–related cyclic development of internal and external spheres of the earth and its space surroundings were considered (Negrutsa and Negrutsa, 1988; Negrutsa 1991, 1994a).

## HIERARCHY OF GEOHISTORICAL STAGES AND CORRESPONDING STRATIGRAPHIC UNITS

As the comparative analysis shows, the structure of the Precambrian sections in the Baltic region is similar to that in other regions of the world (Semikhatov, 1974; Salop, 1982; Borukaev, 1985; Khain and Bozhko, 1988; Basharin, 1991; Negrutsa, 1991; and others). The complete Precambrian sections reveal the common hierarchic structure of no less than nine coordinated ranks in the cyclicity and rhythmicity of geological events (Negrutsa, 1988, 1991; Negrutsa and Negrutsa, 1988). Cycles of the first order—megacycles—record a combination of events that occurred in the time interval between two successive epochs of the continental crust formation or cratonization in terms of A.A. Bogdanov. These epochs, whose boundaries are manifested as the structural–metamorphic unconformities, are marked by major (global) regressions and peneplanation. These events terminate a chain of resonance–connected endogenic and exogenic geoevents, i.e., the complete transgressive–regressive megacycle of sedimentation, magmatism, structural and metamorphic-metasomatic alterations of the earth's crust, which together represent the event content of stratigraphic units of the second order termed as megacyclothem or eonothem (Menner, 1980; Negrutsa, 1988; Semikhatov *et al.*, 1991).[1] The composite section of the stratisphere includes five eonothems: four of the Precambrian (two in the Archean and two in the Proterozoic) and one of the Phanerozoic. Deposits of eonothems are different in composition, structure, stratigraphic and paleogeographic parameters; the deposits of successive eonothems are separated by anisochronous structural-stratigraphic unconformities and past peneplains (Negrutsa, 1988, 1991).

The study on the cyclicity and rhythm of the Lower Proterozoic in Karelia and Kola Peninsula (the third second-rank unit from the base or the Karelian) suggests that the complete megacycle is of the continuous-intermittent transgressive-regressive structure (figure) and accordingly consists of two stratigraphic subdivisions comparable in their time ranges. Its lower and upper units correspond, respectively, to the transgressive or destructive and regressive or constructive branches (megarhythms) of the complete (destructive-constructive) megacycle of sedimentogenesis or tectonic history of earth's crust.

Using the Karelian as an example, it was established in addition that the second-rank units are of a complex

[1] Stratigraphic units of the first order are represented by the Archean and Proterozoic, which are considered as acrothems in the Russian stratigraphic scale (Semikhatov *et al.*, 1991; Semikhatov, 1993).

multimember structure in the regional scale (Negrutsa, 1984; Negrutsa and Negrutsa, 1988). The three-member subdivision of the Karelian and its geohistorical equivalents reflects three principally different geodynamic stages in the evolution of particular regions: (1) early riftogenic, (2) middle platform, and (3) late collisional. Each of them reflects the twofold flooding and draining events, i.e., two complete transgressive–regressive pulses of geosystems or cycles of sedimentation and magamtism. Every pulse is represented by transgressive and regressive deposits, and the corresponding stratigraphic units are subdivided into lower and upper rhythmothems (part A in figure). Both rhythmothems are, in turn, composed of successive transgressive–regressive stratigraphic units (of three ranks at least) and of the corresponding number of coordinated transgressive and/or regressive rhythmothems or rhythmostratigraphic units (part B in figure).

It should be emphasized that the reliable correlation of particular sections directly depends on the rank of the correlated stratigraphic unit, on the detailedness of data on the mutual paleogeographic position of sections, on the reliability and completeness of a certain set of chronologically interchangeable correlative features (and corresponding events), and also on the degree of dislocation and metamorphism of the corresponding supracrustal rock complexes. Therefore, the historical–genetic Precambrian stratigraphy or "cyclorhythmostratigraphy" can, in principle, provide a detailed degree of subdivision of particular sections comparable with that of the Phanerozoic successions. Nonetheless, in practice, these subdivisions are mainly of local and partly of regional importance, and only the high-order units (those of the first, second, or third ranks) retain an interregional and general significance. It can be stated that there is a certain analogy between the modern state of Precambrian stratigraphy and that of Phanerozoic stratigraphy at the time when the biostratigraphic principle of stratigraphic subdivision and correlation was under elaboration.

Thus, cyclo- or rhythmochronostratigraphy furnishes a clue to the comparative geohistorical correlation of Precambrian stratigraphic units with each other and a comparision with Phanerozoic stratigraphic units. Together with the achievements of isotopic geochemistry and paleontology, this allows us to define the correlative interchangeable features (and corresponding events) and to establish individual peculiarities of heterochronous stratigraphic units. Simultaneously, this provides an opportunity to estimate the relative completeness (continuity) of particular stratigraphic successions and to select among them the reference sections and boundary stratotype points of local, regional, and global stratigraphic units. Using this approach, we may obtain in the future the base for the elaboration of the common, in respect of principles and validity, stratigraphic scale of the Precambrian and Phanerozoic (Semikhatov, 1991, 1993, 1995; Semikhatov *et al.*, 1991; Negrutsa, 1992, 1994b; Prozorovskii, 1995).

## GEOEVENT METROLOGY

At the present level of knowledge, the Precambian, embracing about 3.5–3.6 m.y. of the earth's history, can be subdivided into four global subdivisions which correspond to the unite of the second rank: (1) Lower Archean (Eoarchean), comprising rocks that form the basement of structural–lithological associations in the Archean greenstone levels (older than 3600–3800 Ma); (2) Upper Archean, corresponding to rocks older than 2800–2500 Ma; (3) Lower Proterozoic (Karelian); and (4) Upper Proterozoic (Riphean). The boundary between the last two stratigraphic units is dated at 1650 Ma. As known, all these subdivisions are characterized by a different composition and structure and substantially differ from each other. Simultaneously, they reveal the common transgressive–regressive pattern and characterize, as shown by the well-known data of isotopic chronometry (Salop, 1982; Semikhatov, 1993, 1994; and others), an almost equal duration (about 1000 m. y.) close to the complete period of rotation of the solar system around the Galaxy center (1056 m. y., according to Kulinkovich, 1992). Correspondingly, the duration of the transgressive and regressive halves of the subdivisions may be close to 500 (528) m. y.

The Upper Proterozoic (Riphean) unites three empirically separated stratigraphic units: Burzyanian, Yurmatinian, and Karatavian. All of them are, as shown by data of isotopic geochronology, of the approximately equal duration close to one third of the rotation period of the solar system around the Galaxy center (352 m. y.). As follows from the aforesaid, the similar three-member subdivision is also peculiar of an integral Karelian (Lower Proterozoic) section in the stratotype area and in stratigraphic equivalents of the section in other regions of the world (Salop, 1982; Negrutsa, 1991), and the available isotopic–geochronologic data suggest the same duration of these units. It is hardly incidental that each of six subdivisions of the Proterozoic appears to be similar in duration to the Paleozoic and, correspondingly, to one third of the rotation period of the solar system around the Galaxy center.

Similarly to the Proterozoic stratigraphic units comparable in rank and duration, the Paleozoic reflects events of two complete transgressive–regressive and corresponding tectonic–magmatic (Caledonian and Hercynian) cycles, whose duration is close to $170 \pm 10$ (176) m. y. As known, the Mesozoic is of a similar continuation. Thus, the time range of three Phanerozoic cycles together is about 500 m. y., i.e., equal to the half of each complete (1000 m. y.) Precambrian megacycles. It follows from the aforesaid that the stratigraphic units of equal duration also correspond to each other in the geohistorical content. All this is consistent with the concept suggesting the uniform causes and regular pulsatory development of the geospheres during the observable history of the earth.

If this is true, then one can assume that similar characteristics are generally valid also for the cyclo- and

rhythmostratigraphic units of lesser coordinated ranks, and thus the principle of periodization of the geological history based on natural events provides the stratigraphy with a chronological standard for the assessment of approximate duration of stratigraphic units physically expressed but lacking concrete chronometric characteristics and of stratigraphic hiatuses. All this allows me to admit that the search for a theoretical model of metrology of the geological time recorded in rock successions (Afanas'ev, 1987, 1991; Zakoldaev, 1992; Barenbaum, 1995) and the periodical law of the historical geology of Kulinkovich (1985) are both realistic. Some uncertainty in the isotopic–geochronological dating of boundaries and formation of geological bodies reflects the lateral diachronism of corresponding geological events. The scale of diachronism for boundaries between structural and lithological bodies is in direct proportion to the duration of geohistorical stages, which are characterized by these bodies.

## CONCLUSION

The comparative analysis of periodicity in the interrelated evolution of Precambrian geospheres that are almost completely materialized in deposits of marine transgressions and regressions allows the conclusion that the geoevent approach to the study of geological history provides a mean for the unity and comparable degree of detail of the historical–genetic division of the Phanerozoic and Precambrian into periods. In other words, this approach is more universal than the biostratigraphic one. Simultaneously, the geoevent approach more distinctly differentiates the peculiar features of the Precambrian and Phanerozoic.

In the context of this paper, an important point is that the earth experienced four complete megacycles during its Precambrian history and at the later stage of its geological history entered the second half of the fifth, still continuing the Phanerozoic megacycle. Accordingly, the modern state of the earth is comparable with that characteristic of the initial stage of the second (regressive) megarhythm of the complete megacycle. Therefore, it seems to be correct to reason that the actualist approach to decoding the Precambian earth history can result in an adequate interpretation only when applied to deposits of brief periods corresponding to the regressive branches of Precambrian megacycles terminated, according to calculations, at about 1050, 2100, and 3150 Ma ago. The respective stages coincided in time with changes in the cyclically developing mantle dominant that affected the geochemical background of the outer geospheres and specific crustal processes (Negrutsa, 1984, 1991).

It is clear that these assumptions give rise to a series of fundamental consequences such as, for instance, the wavelike evolution of the organic world, the resonance–multiaspect interrelation between life and the environments, etc. (Negrutsa, 1994a). The effectivity of our understanding of periodicity of geoevents in the geological history corresponds to modern achievements in the study of geological cycles, in deciphering the relationships between the earth and space, and in widening the stratigraphic methodology. It is apparent that such investigations, which were stimulated by works of A.P. Karpinskii and further developed by B.L. Lichkov, V.V. Menner, and other researchers, are now largely ignored by the practice of historical geological and, moreover, by biostratigraphy and geochronology. Meanwhile, the comparative cyclorhythmology promises principally new information important to natural science as a whole, not solely to earth science (Negrutsa and Negrutsa, 1996).

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# Comments on the Article "The Late Stage of Evolution of the Northern Part of the Okhotsk–Chukchi Volcanogenic Belt" by V. F. Belyi, M. I. Raikevich, and B. V. Belaya[1]

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In connection with the publication of Belyi and coauthors, we consider the problem of priority in scientific investigations; paleobotanical records, which may cause the establishing of an older age of the continental deposits of the Sea of Okhotsk–Chukchi volcanogenic belt (OCVB); and the consequences of some misunderstandings for the geological science and practice.

The problem discussed in the article of Belyi *et al.* "The Late Stage of Evolution of the Northern Part of the Okhotsk–Chukchi Volcanogenic Belt" is evidently urgent, because the determination of the upper age limit of the belt is of great importance for scientific and practical problems concerning the geology, structure, and evolution of the OCVB and adjacent regions, as well as the role of the belt formations different in age in controlling the ore genesis.

The authors of this article used the paleophytological and palynological methods to establish the Senonian (Santonian) age of the late stage in evolution of the central OCVB sector, whereas the individual units of the studied sections are correlated on the basis of paleomaghetic data only. Belyi (1994) inferred some time ago the mid-Albian–Santonian age of the OCVB as a whole. However, there is no reference in either articles to the numerous publications of the last two decades, in which the Senonian age of the late stage of the OCVB was substantiated. We also cannot agree with the authors' statement that their article presents the first successful results of paleophytological and palynological investigations, because the palynological evidence for the Senonian age of the upper part of the OCVB volcanic sequence were first reported more than 15 years ago (Bityutskaya *et al.*, 1979).

That is why we have to briefly consider the history of views on the age of the OCVB volcanogenic formations.

The Interdepartmental Regional Conference on Stratigraphy of 1957 held in Magadan attributed the volcanogenic formations of the northern part of the belt (the Chaun Group) to the undivided Upper Cretaceous and the Okhotsk and Evenki groups of its southern part to the Lower and Upper Cretaceous, respectively. Later (*Geologiya ...*, 1970), the Aptian–Danian (Lower and Upper Cretaceous) and Paleogene deposits were recognized within the OCVB.

On the basis of abundant fossil plants of the Cretaceous age from the northern part of the belt, Belyi (1969) referred the OCVB volcanogenic formations to the Valanginian–Danian. However, somewhat later, Belyi changed this age interpretation and limited the stratigraphic range of the OCVB to the Albian–Cenomanian (Belyi, 1975, 1977, 1981, 1988, 1992, and others). In his numerous publications of the last two decades, he insisted on the Cenomanian (even early Cenomanian) age of volcanic rocks from the upper part of the OCVB section. At the same time, he suggested the same age for the plateau basalts overlying both the OCVB formations and the Upper Cretaceous molasses of the adjacent Anadyrka–Koryak region. Both the structural position and petrochemical composition of the plateau basalts show that they do not belong to the volcanic belt of the continental margin (Filatova, 1987, 1988, 1995). The age, composition, and geodynamic setting of the plateau basalts of the interplate type are subjects of special consideration, so we omit this issue here.

Belyi's opinion to accept the Senonian as the upper age limit of the OCVB was approved at the 2nd Interdepartmental Regional Conference on Stratigraphy in Magadan in 1975 (*Resheniya ...*, 1988) and used in many different-scale maps (geological, tectonic, structural–lithological, and others) compiled by teams of Geological Survey for the Northeastern Territories.

However, as far back as the 1960s and 1970s, floral remains of the Turonian and Senonian ages were found in the OCVB and adjacent areas by geologists of the survey and AO Aerogeologiya, who carried out the middle-scale geological mapping and special biostrati-

[1] Stratigr. Geol. Korrelyatsiya, 1997, vol. 5, no. 5, pp. 78–89.

graphic studies there. The most representative collections of the Late Cretaceous flora were made by L.A. Ankudinov, N.S. Afanas'eva, A.I. Dvoryankin, A.D. Devyatilova, V.A. Gretskii, V.V. Gromov, V.A. Zakharov, A.I. Milekhin, E.B. Nevretdinov, G.P. Terekhova, O.P. Timofeev, Yu.S. Turchinovich, V.A. Faradzhev, N.I. Filatova, G.G. Filippova, Yu.M. Khotin, A.B. Tsukernik, V.I. Chubarov, and others.

The occurrence of the Turonian and Senonian formations in the OCVB section has been reported in a series of publications (Filippova, 1975; Devyatilova *et al.*, 1980; Nevretdinov, 1980; Terekhova and Filippova, 1983, 1984; Samylina, 1984a, 1984b; and others).

The authors of these comments were also involved into the study of the OCVB region. Filatova, who was first researcher at the AO Aerogeologiya and then at the Institute of Lithosphere, Russian Academy of Sciences, conducted special stratigraphic investigations in 1968–1991 on the basis of the large-scale geological mapping. Lebedev, researcher at the Geological Institute, Russian Academy of Sciences, has been involved in special phytostratigraphic studies since 1973. The joint investigations in the OCVB region extending from the Ul'ya sector in the southwest to the central sector of the Chukchi Peninsula in the northeast provided us with rich collections of fossil plants. The newly distinguished floral assemblages and previously published data allowed us to better evaluate the age of the belt and to define a specific character of its Cretaceous floras (Lebedev, 1975, 1979, 1987, 1992; Filatova, 1988; Filatova *et al.*, 1977; Filatova and Lebedev, 1982; Lebedev *et al.*, 1989; Dvoryankin *et al.*, 1990; and others). The terrigenous–volcanogenic OCVB section was found to be characterized by eight floral assemblages widespread over the whole OCVB and spanning the interval of the upper Albian to Campanian.

The most important results of the joint investigations were the finding of the Senonian floral and palynological assemblages in the upper part of the OCVB section and the establishing of the Campanian age of its uppermost formations. The Senonian formations dated by means of fossil plants were recognized throughout the OCVB, from its southwestern part to the central sector of the Chukchi Peninsula (Filatova and Lebedev, 1982). The Campanian age was supported by the occurrences of the Ust'-Emuneret flora (the Central Chukchi sector), the Murgal and Ayanka floras (the Penzhina sector in the central part of the OCVB), and the Delokachan flora (the Ul'ya sector). The Tunum flora of the early Senonian age was found in the southwestmost part of the OCVB.

It is significant that the occurrence of the Senonian deposits in the OCVB was confirmed at the same time by palynological investigations. In addition to pollen assemblages of Albian–Turonian age, an early Senonian assemblage was identified in the Ul'ya sector (Bityutskaya *et al.*, 1979). G.D. Davydova and G.M. Brattseva distinguished an early Senonian palynoassemblage of similar composition in the Penzhina sector of the OCVB. These assemblages contain pollen of the *Triproectacites* and other angiosperm groups. It should be stressed that both assemblages were tentatively dated as early Cenomanian, but they can be referred to the younger Senonian level due to the presence of *Triporopollenites plicoides*, *Loranthacites*, *Manicorpus*, and *Orbicularis* in the spore–pollen spectra (Bityutskaya *et al.*, 1979).

Thus, the results of our biostratigraphic and mapping studies during the last two decades show that the OCVB development spans the Albian–Campanian time interval (not less than 30 m.y.), which is two times longer than the Albian–early Cenomanian interval suggested by Belyi (1977, 1982, 1988, and others).

In his later sharp review, Belyi (1992) continued to insist on the Albian–early Cenomanian age of the OCVB and denied the Campanian age of the Delokachan flora from the Ul'ya sector (we correlate it with the Ust'-Emuneret flora) considering it as correlative to the Grebenka Formation or, to be exact, to the Dukchanda floral assemblage of the Cenomanian age.

However, shortly after, at the international meeting in South Korea (1994), Belyi reported on the Senonian age of the upper part of the OCVB section without any references to our studies (Belyi *et al.*, 1994).

Another publication (Belyi, 1994) contains the analogous information also without mentioning our studies. In the article under discussion (Belyi *et al.*, 1997), our works were cited only to state that the previous Senonian datings to be incorrect in contrast to the "true" Senonian formations of the OCVB discovered by Belyi and his colleagues.

In the second half of our discussion, we shall briefly touch on some problems of paleobotany and phytostratigraphy, which are of principal importance for estimating the development time of the OCVB.

The older age for the Late Cretaceous floras was inferred at the initial steps of paleobotanical studies in the OCVB, when a specific character of its floras evolved in environments of the terrestrial volcanicsm. The OCVB floras include, at least till the Campanian level, the abundant Early Cretaceous genera *Phoenicopsis* and *Sphenobaiera*, which became extinct in the coastal environments at the beginning of the Late Cretaceous. Many coastal floras correlative with marine faunas served as the guide fossils for dating the terrestrial volcanogenic deposits. The automatical consideration of specific features of the coastal floras (in particular, the absence of *Phoenicopsis* and *Sphenobaiera* in deposits younger than Cenomanian beds) as characteristic of the Late Cretaceous OCVB floras resulted in speculations about their older (up to the Cenomanian) age and, hence, in the incorrect dating of the formations.

The OCVB is a part of the North Pacific floral refugium distinguished by Vakhrameev (1981). Some mesophytic gymnosperms, cycadophytes, in particular,

still existed in the refugium during the Late Cretaceous, when they became extinct in other areas of the globe. In the OCVB, there were *Phoenicopsis* and other forms, which had virtually disappeared by that time in plains. Therefore, some coeval Cretaceous floras from the terrestrial volcanogenic and coastal–marine sedimentary deposits of northeastern Asia may be quite different in composition. The specific character of the Late Cretaceous floras of the OCVB allowed this region to be recognized as an independent paleofloral subprovince (Lebedev, 1987, 1992).

*Phoenicopsis* and *Sphenpbaiera* cannot be considered as leading forms of the Late Cretaceous floras of the OCVB and indicators of their Cenomanian age. In the Late Cretaceous, the guide fossils characteristic of certain phytostratigraphic horizons were, as a rule, angiosperms, which most rapidly evolved at that time. To avoid misunderstanding, the collections of the OCVB fossil plants lacking angiosperms should be assigned to the undivided Cretaceous or to its upper series.

For instance, the Ust'-Emuneret floras were dated on the basis of abundant remains of *Quercus tchucotica* and associated specimens of *Macclintockta ochotica* (Abramova, 1979; Herman and Lebedev, 1991). Both angiosperm species, which appear in the section beginning from the Barykov Formation (the lower Campanian), are missing in sediments below this level (Lebedev, 1987, 1992). Taking this into account, we cannot agree with Belyi, who attributes the Ust'-Emuneret flora to older deposits because "...the Barykov flora... contains no mesophyte relics, such as *Sphenobaiera*... and *Phoenicopsis*..., which are abundant in the Ust'-Emuneret assemblage" (Belyi, 1997, p. 86). As shown above, the usage of the Early Cretaceous forms for dating some OCVB floras may cause (and caused in the last three decades) some significant stratigraphic mistakes.

Owing to the specific character of the OCVB floras, reflecting the peculiar paleogeographic environments, numerous attempts to correlate them with floras from other, even proximal, paleogeographic areas seem to be doubtful. This especially concerns the Arkagala flora of the Verkhoyansk–Chukchi tectonic area. The latter characterizes the swampy lowland conditions (Samylina, 1988) quite different from those of the OCVB. That is why it is evidently impossible to discover in the OCVB formations the plant assemblage completely identical to the Arkagala type flora. Therefore, the term "Arkagala flora" is inapplicable to the OCVB floras. In our opinion, the only correct approach is to distinguish the OCVB floras proper and to correlate them to the floras of adjacent regions taking into account the specific paleogeography of the latter.

In conclusion, it should be noted that it does not matter who was the first to prove the Senonian age of the last stage of the OCVB evolution. It is more significant that, in spite of the available paleobotanical and palynological data, there are attempts even now to constrain the stratigraphic range of the OCVB sections subdivided in detail by geologists of the Geological Survey for the Northeastern Territories to the Albian–Cenomanian interval, i.e., to make them two times shorter. Naturally, this negatively affects the geological practice (metalogenic analysis and prognosis) and science.

In addition, the reduction of the time interval of the OCVB development resulted in some contradictory attempts to correlate the belt formations with the coastal–marine deposits of the Koryak region and to reconstruct the evolutionary stages of the Asian margin during the Cretaceous on the basis of the plate-tectonic concept (Belyi, 1977; Belyi and Pokhialainen, 1990; Filatova, 1988, 1995; and others).

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Reviewres M.A. Akhmet'ev and A.B. Herman

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# Reply to Comments on the Article "The Late Stage of Evolution of the Northern Part of the Okhotsk–Chukchi Volcanogenic Belt" by V. F. Belyi, M. I. Raikevich, and B. V. Belaya[1]

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**Key words**: *Okhotsk–Chukchi volcanogenic belt, late stage, Turonian, Coniacian, Santonian, palynoflora, magnetostratigraphy*

From the very beginning to the last page, the article under discussion is devoted to new data on stratigraphy, palynology, magnetostratigraphy, and paleovolcanology of the area of the Enmyvaam River upper courses studied during field work in 1991 and 1993. This area exhibits the most complete section of volcanogenic formations of the late stage in evolution of the Okhotsk–Chukchi volcanogenic belt. The materials obtained are of principal significance for substantiating the upper age limit of the volcanogenic belt. Moreover, the methodological importance of these investigations is beyond doubt, because they represent the first successful attempt of the comprehensive study on the Cretaceous continental stratigraphy of not only the volcanogenic belt, but the entire northeastern Asia as well. The rocks of the reversed polarity R-zone were discovered for the first time in the upper part of the Middle Cretaceous continental deposits. On the basis of a combination of different (palynological, stratigraphic, paleovolcanological) features, these rocks turned out to be correlative with the lower marine deposits of the Barykov Formation corresponding to the Klyuevo R-zone, and *Inoceramus involutus* Zone of the lower Coniacian (Pergament *et al.*, 1971; *Paleomagnitologiya*, 1982).

The article does not concern general problems, such as the priority of recognition and substantiation of the Senonian deposits of the volcanogenic belt sections and, especially, the disputable problem of dating of the entire volcanogenic belt–suggested as the main topics of the discussion by E.L. Lebedev and N.I. Filatova. Therefore, leaving aside these general problems, I shall dwell on the concrete results of our work that do not agree with the viewpoints of our opponents.

(1) The article shows that the stratigraphic units of the late developmental stage of the volcanogenic belt correspond to the succession of the Ergyvaam–Emuneret–Enmyvaam formations, but not that of the Emuneret–Ergyvaam–Enmyvaam formations (*Resheniya* ..., 1978). Thus, the Ust'-Emyneret flora is sampled from the lower part of the Emuneret Formation, but not from the Ergyvaam Formation, as Filatova suggests (Filatova and Lebedev, 1982; Filatova, 1988).

(2) We recovered the representative pollen assemblage of the late Turonian age from the same beds (the same trenches, stripped areas, and samples), where Lebedev collected the Ust'-Emuneret flora. These beds occur in the lower part of the Emuneret Formation displaying the normal magnetization. The middle part of this formation with the reverse polarity is correlated with the Klyuevo R-zone of the lower Coniacian (Pergament *et al.*, 1971; *Paleomagnitologiya*, 1982). Therefore, we consider the Ust'-Emuneret flora to be of late Turonian but not Campanian age, as suggested by Lebedev only based on the occurrence of *Quercus tchucotica* Abram.

(3) The Enmyvaam Formation belongs to the sequence of high-Al plateau basalts (Belyi, 1994), which terminate everywhere the section of the Okhotsk–Chukchi volcanogenic belt. In the Enmyvaam Formation, we identified a representative pollen assemblage of late Turonian–early Senonian age. The Enmyvaam Formation was considered as the Santonian in age on the basis of its stratigraphic position (above the Klyuevo R-zone) and normal magnetization of the rocks. Neither Lebedev nor Filatova can accept this date, because they attribute the plateau basalts to the Upper Cretaceous–Paleocene or Danian (Lebedev, 1987; Filatova, 1988; Grigoryants *et al.*, 1992) contrary to the earlier published data on the Middle Cretaceous age of the plateau basalts exposed at the Gipoteticheskii Creek in the Okhotsk sector of the volcanogenic belt (Belyi and Samylin, 1987).

It should be noted in conclusion that Lebedev and Filatova erroneously believe that I previously dated the beginning of evolution of the Okhotsk–Chukchi volcanogenic belt back to the Valanginian. In my first thorough publication on the geology of the belt (Belyi,

[1] *Stratigr. Geol. Korrelyatsiya*, 1997, vol. 5, no. 5, pp. 78–89.

1961), this stage was referred to the Aptian, and then I attributed it to the Albian.

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